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FACTS AND HYPOTHESES CONCERNING THE GENESIS OF DOLOMITE ROCKS

by

N. M. Strakhov

The genesis of dolomite rocks is generally recognized as one of the most difficult in theoretical sedimentary petrology. Even today it is possible in this subject to come across many contradictory opinions which tend to make things difficult, especially for the beginning research worker in this field. During the past ten years, however, factual material has been accumulated by degrees; on objective analysis this material narrows the limits of theoretical divergence and directs research into a more definite channel. Unfortunately, many research workers are not as yet sufficiently conscious of the new facts which have emerged and do not study them fully. The task of this report is to give a brief resume of the modern, factual side of the problem as I see it, and to present a genetic theory which is prompted by these facts and is sufficiently in accord with them.

* * * * *

I. Basic Factors Affecting the Genesis of Dolomite Rocks.

Six cardinal geologic, and in part, geoghemical factors must be kept in mind in studying the problems of the genesis of dolomite rocks.

1. The first factor is the multiple character of dolomite rocks, i.e., their genesis in different environments [6]. At the present time lake dolomites are recognized which may be divided into at least two types; sodium (Tanatar lakes) and magnesium carbonate (Balkhash and the lakes of the Burlinsky group). We also have lagoon dolomites -- both salt- and fresh-water, and finally, marine dolomites which occur both in regional, semi-isolated areas and in the central parts of marine basins. The multifacial characteristic of dolomite accumulations is visually represented on the diagram (Fig. 1) on which the first row summarizes the data concerning the distribution of dolomite in modern basins; the second row -- in Paleozoic basins.

2. Notwithstanding the diversity in origin of facies of dolomite, there are in all only two petrographic types of dolomite rocks -- bedded and metasomatic.

In bedded rocks the extension of layers over great areas is characteristic; they have been measured not merely in kilometers but in tens and even hundreds of kilometers as, for example, the Kashir dolomites (I. V. Khvorova), and the dolomite beds with "worm tracks" on the border of the Steshev and Protvinsk beds in the northwest part of the Moscow syncline (S. G. Vishnyakov).

The composition of the rock corresponds to normal dolomite, completely devoid of calcite, or containing only small percentages of $CaCO_3$ in the form of a solid solution. The structure is always non-lithomorphic or micro-grained and fine-grained with very weakly-defined indications of recrystallization. Organic remains are absent, or present only rarely; for the most part these are only ostracods and more rarely brachiopods of a distinctive type. The faunas on the whole form a distinctive assemblage, clearly distinguished from those of the Carboniferous above and below them.

In the dolomite layers there commonly are traces of lamellar crystals of gypsum; after leaching only crevices containing them remain, and instead gypsiferous fluorite is sometimes present (e.g., in the Kashir dolomites). Bedded dolomites in places contain quantities of a fine pelitic material

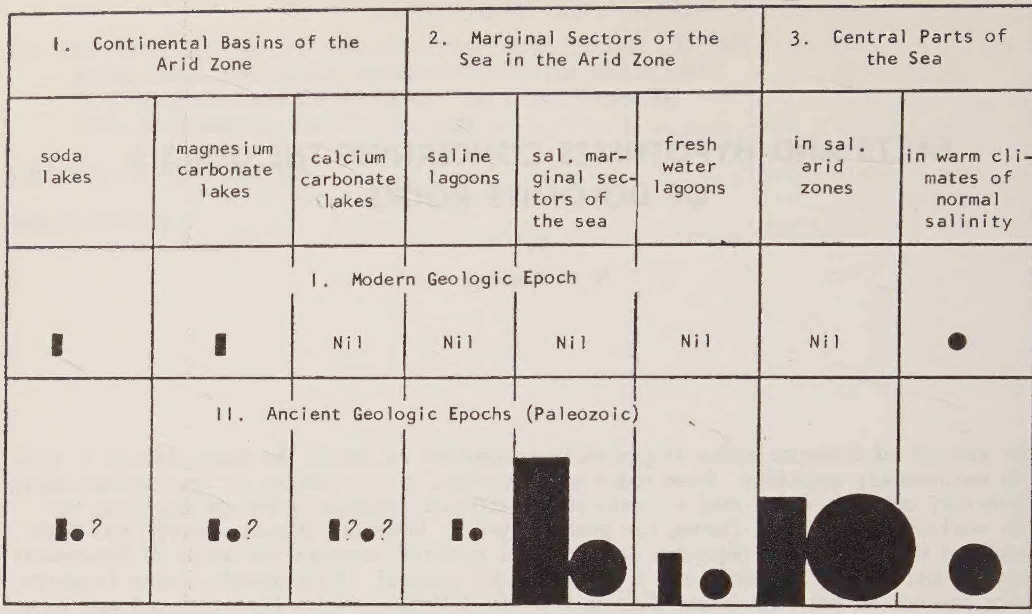

FIGURE 1. Facies conditions in the formation of dolomite.

Bedded dolomite (sedimentary) is represented by the columns; spotted metasomatic dolomite (sedimentary-diagenetic) by the circles.

which makes them argillaceous almost to the point of being dolomite marls, and gives rise to the beginnings of cleavage. Among the argillaceous minerals montmorillonite is common, also palygorskite and sepiolite. The most important diagnostic indication of bedded dolomite is the absence of perceptible traces, however small, of the metasomatic replacement of the earlier metasomatic replacement of the earlier accumulations of calcium carbonate by dolomite. Only where there is present in the dolomite bed cretaceous material wholly or partly filled with dolomite, is it possible to speak of the metasomatism of calcite to dolomite, but the scale of this transformation in any given instance is, understandably, quite negligible. Metasomatic dolomites are distinguished by an extreme diversity in forms of occurrence; they sometimes occur in bedded formations of complex configuration kilometers and even tens of kilometers in extent; in other instances they are developed in fantastic designs and complicated lens forms hundreds and tens of meters in extent and often broken down into separate shreds. Farther on, there may be stocks of varied contour, meters in length and magnitude, and lastly tiny patches and veins tens of centimeters in extent. In all cases, however, three typical characteristics distinguish this type of dolomite rocks from bedded dolomites. One of these is the unusual variation in the degree of dolomitization within the seams, lenses, and spots (even in the bedded formations). The dolomite content can vary from 95 percent to 2 percent and less; and the localization of the areas of high and low dolomite content is extremely capricious and is not subject to any clearly defined law. Furthermore, indication of the metasomatism of dolomite from calcite are easily traceable in the structure of the dolomite seams and patches.

The earliest stage of all contains fine-grained dolomite grains in which are developed both regular rhombohedra, or clusters of them, and areas of dolomite grains with irregular outlines. Later, the rock contains remains of Crustacea, Foraminifera, Brachiopoda, Pelecypoda, etc., up to Crinoidea which become dolomite in the final stage. Generally speaking, the more coarse-grained the replaced calcite (inorganic or biogenic in origin), the more difficult is the process of its replacement by dolomite, and the later the stage to which replacement itself is postponed. There remains one final characteristic of this type of metasomatic dolomite rocks; the absence in them of their own particular type of fossil assemblage. We may add that dolomites of the second type are commonly porous, cavernous, and

friable, and frequently bear traces of the addition of iron, which expresses itself in a reddish or rust-brown hue of varying intensity. In regard to the distribution of the various petrographic types of dolomite in relation to the precipitation of the different facies, modern developments in this aspect compel one to hold the opinion that both types of dolomite rocks occur under all circumstances.

3. An analysis of hydro-chemical conditions in modern and ancient basins, both lake and marine, in which dolomite deposits have emerged, shows that in the vast majority of cases it is possible to observe clear signs of some salinification of the basin; and in a number of them an increase in the alkaline content and pH. In this connection it is noticeable that the formation of minerals follows one and the same sequence; first calcite and later -- with an increase in the salt content -- dolomite.

Thus, in modern soda lakes with a concentration of up to 0.1 percent, only calcite is precipitated. In lakes of higher concentration, dolomite manifests itself by comprising up to 40 percent of the sum of the carbonates where S=0.2 to 0.3 percent and from 90 to 100 percent where S=1 percent (Table 1) [3].

In modern lakes containing carbonated water (up to a salt content of 0.2 percent) at first only calcite is formed pure or with a negligible admixture of dolomite (traces up to 1.5 percent). In some places this calcite is organic, often to quite an appreciable degree. With a salt content of 0.5 to 1.5 percent, the dolomite content in the sediment rises to 60 to 70 percent, and with a content of 4.5 percent, up to 90 percent of the sum total of the carbonates is dolomite. When the salt content becomes very high -- over 14 percent -- the dolomite almost completely disappears, being replaced by separate deposits of calcite and magnesite.

There are no traces of dolomite in modern lakes, but they are extremely characteristic of many Paleozoic sea basins: e.g., D_3 = C - P of the Russian Platform; Cm - S of the Siberian; they also occur on the Chinese and North American platforms and in many others. In these cases, the dolomite rocks are located in the form of huge accumulations in some instances on the periphery of a sea, in others in its central regions. In the latter cases, the border zones have no dolomite or a very low content, and are composed of calcareous rocks. This dual localization in the formation of dolomite is easily visible on the maps and diagrams of A.B. Ronov for the Upper Paleozoic of the Russian Platform (Figs. 2 to 5) and on that of K.K. Zelenov for Cm_1 of the Siberian Platform [6] (Fig. 6).

Some of the saline portions of the border dolomite formations are marked by the presence of fluorite in the dolomite strata, occasionally by beds and whole horizons of gypsum, and lastly by the progressive impoverishment of the organic fauna in proportion to the development of dolomite-bearing rocks. However, the dolomites which appeared in the central regions of the Carboniferous seas of the Russian Platform show clear indications that they were formed in parts of the sea having a somewhat higher salt content than the peripheral zones where there was no accumulation of dolomite [6]. The distribution of the fauna in the carbonate strata bears witness to this. In the calcareous precipitates along the periphery of the seas, there is an abundant fauna of all kinds, and it is from the study of these fauna that modern ideas about the biota of Carboniferous seas originated. In dolomite rocks, organic remains are far more meager and the faunas are uniform. From this it may be seen that the impoverishment of organic life, at least in a number of instances, is not a secondary manifestation of the formation of metasomatic dolomites, but a fundamental peculiarity of them. The Upper Carboniferous of Samarskiye Luki serves as a classic example of this, and even forty years ago M.E. Noinskiy pointed out that the upwards impoverishment of the fauna, which was paralleled by the development of dolomite rocks, was a decided indication of the progressive salinification of the sea. In exactly the same way it is generally accepted that the impoverisation of faunas in the Upper Famennian stage paralleled by the enrichment of the deposits with dolomite was a similar, primary feature. The analogous correlations by A.B. Ronov are designated C_1 for the summits and the Middle Carboniferous of the Central Russian Platform.

Thus, we have a basis on which to form opinion that the central regions of the Upper Paleozoic of the Russian Platform having accumulated dolomites were actually distinguished from their peripheral littoral zones by a somewhat higher salt content. It is indeed difficult to estimate the index of salinity, but taking into consideration that in the modern Red and Mediterranean Seas the rise in salinity to 38 to 42 percent, i.e., 20 percent above the normal, has not as yet caused the suppression of the marine fauna to any appreciable extent; on the other hand, in the dolomite regions of ancient seas the fauna was suppressed. Consequently, the salinity of the Upper Paleozoic basins of the Russian Platform, in separate eras at any rate, exceeded the peripheral (normal) by 8 to 20 percent and possibly even more, and

Table 1

The carbonate composition in Modern Lake deposits

Basin	Calcite percent of sediment	Dolomite percent of sediment	Magnesite percent of sediment	Percent dolomiticity of carbon	Salinity of Basin
I. Soda Lakes					
Rublevo	34.47	22.31	None	38.5	
Tanatar IV	6.84	31.54	"	82.0	
Tanatar V	3.3	23.9	"	87.0	
Tanatar III	0.13	14.0	"	100.0	
II. Magnesium Carbonate Lakes					
Maloye Topol'-noye	14.48	None	None	0	
Khomutinoye	57.50	"	"	0	Fresh to 0.2%
Peschinoye	25.68	"	"	0	
Balkhash, western reaches	23.41	1.45	"		
Bol'shoye Topol'noye	27.24	6.54	"	19	
Balkash, eastern reaches	28.43	37.75	"	56	
Balkash, eastern reaches	29.64	32.37	"	52	
Balkash, eastern reaches	21.52	33.88	"	61	0.2-4.5%
Bol'shoye Kulundinskoye	3.8	18.25	"	83	
Bol'shoye Kulundinskoye	10.83	27.86	"	72	
Bol'shoye Kulundinskoye	3.01	20.06	"	90	
Anzhebulat, white mud	7.03	None	30.95	--	
Anzhebulat, white mud	5.39	"	10.15	--	
Near-Balkash Lakes					Brine of a higher salinity than 14%.
Insoluble thenardite fragments	12.38	"	49.54	--	
Insoluble thenardite fragments	21.29	"	24.17	--	
Insoluble thenardite fragments	15.29	"	41.91	--	

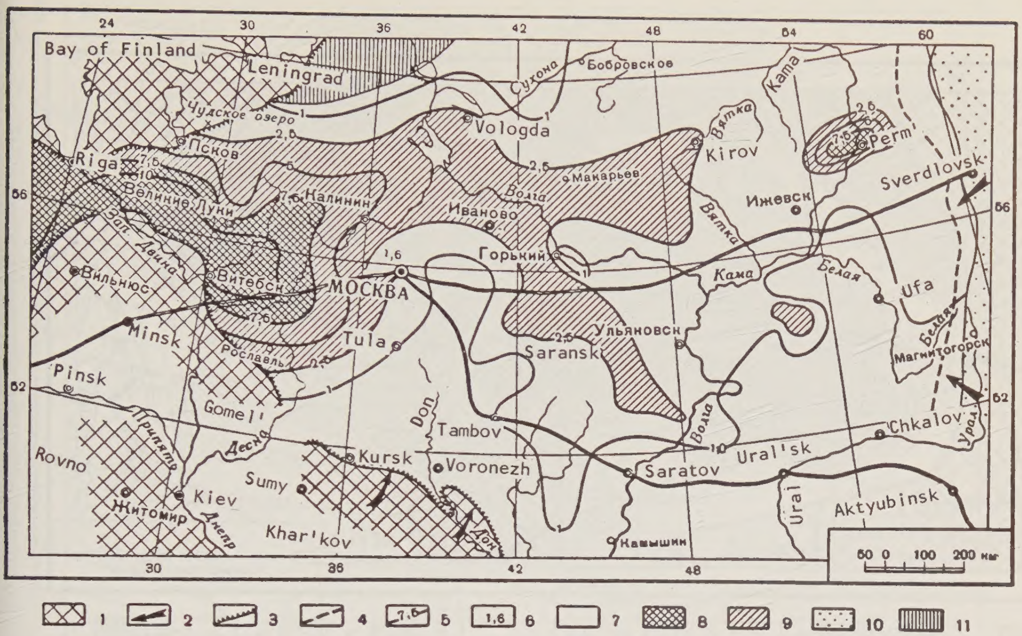

FIGURE 2. The Distribution of Magnesium in the Carboniferous Rocks of the Frank Stage of the Russian Platform. (according to A. B. Ronov, simplified)

1-Regions of Erosion; 2-Direction of the Transference of Fragmental Material and the Probable Courses of the Discharge of Fresh Water; 3-Boundary of Regions of Erosion; 4-Boundary of Russian Platform and Urals Geosyncline; 5-Lines of Equal Mg Content (in %); 6-Average Mg Content in a Layer (in %); 7-Regions of Predominant Occurrence of Limestone; 8-Regions or Predominant Occurrence of Dolomite; 9-Regions of Occurrence of Carboniferous Rocks, Intermediate between Limestone and Dolomite; 10-Littoral Marine Terrigenous Deposits; 11-Continental Deposits.

was 1-1/2 to 2 times higher than that of the peripheral.

The discovery of dispersed crystals of fluorite in bedded dolomite, and of occasional localized seams of gypsum serves as a supplementary proof of this. The existence of a somewhat higher salt content in the central dolomite-forming zones of the Carboniferous seas is easily explained by the increased evaporation of water from the upper basins in arid climatic conditions. In the littoral zones, this increase in evaporation was compensated by an undercurrent of fresh water emanating from the continents; in the central portions of the basins these currents were practically never effective.

Thus, regardless of whether the dolomites formed in lakes or lagoons, in the marginal isolated zones of the ancient sea basins or in the central portions of the platform seas, their formation was invariably characterized by some increase in the salt content of the water compared to those waters where dolomite did not form.

This state of affairs indicates with certainty that massive accumulation of dolomite in ancient lakes and seas was indigenous, not to all climatic zones, but to quite clearly defined ones. Beginning in the Lower Paleozoic the accumulation of massive dolomite formations was a process characteristic of lithogenesis in the arid zone. In humid zones it was confined almost exclusively to reef facies.

4. The formation of dolomite was clearly defined and regulated during salinification of the basins of the arid zones. This view was initially defined by me in cooperation with A.I. Tsvetkov in the example of the Kungursky precipitations of the Southern Pre-Urals. As may be seen on the diagram (Fig. 7), the dolomite formation came after the calcite and achieved its maximum extent during the deposition of gypsum.

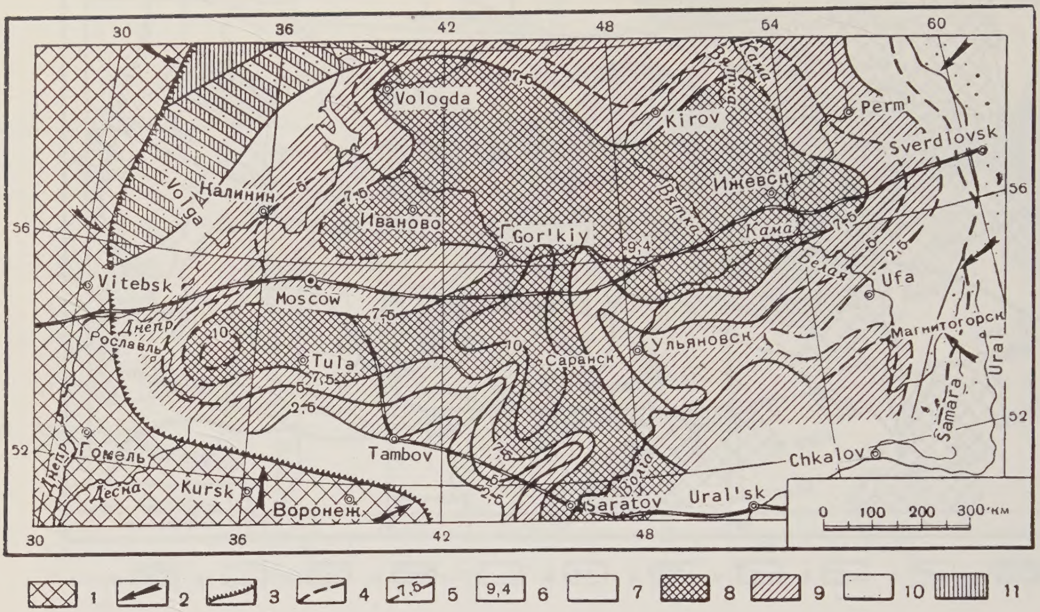

FIGURE 3. The Distribution of Magnesium in the Carboniferous Rocks of the Famennian Stage of the Russian Platform. (according to A. B. Ronov, simplified)

1-Regions of Erosion; 2-Direction of the Transference of Fragmental Material and the Probable Courses of the Discharge of Fresh Water; 3-Boundary of Regions of Erosion; 4-Boundary of Russian Platform and Urals Geosyncline; 5-Lines of Equal Mg Content (in %); 6-Average Mg Content in a Layer (in %); 7-Regions of Predominant Occurrence of Limestone; 8-Regions of Predominant Occurrence of Dolomite; 9-Regions of Occurrence of Carboniferous Rocks, Intermediate between Limestone and Dolomite; 10-Littoral Marine Terrigenous Deposits; 11-Continental Deposits.

Subsequently, deposition of dolomite was sharply curtailed, and was replaced by a distinct precipitation of calcite and magnesite when the salt content increased to that required for the deposition of halite. In the Paleozoic lagoons of marine origin, such an evolution of carbonate beds may be universally observed without exception. In the Mesozoic and Tertiary periods, the formation of dolomite as a consequence of the salinification of the sea lagoons is less certain; in places it is present, in places not. Thus, for example, with the salinification of the basins of the Upper Jurassic of Germany and the Caucasus, dolomite strata were formed; on the development of the contemporaneous lagoons of South Tadzhikistan the dolomite phase came to an end and limestones were directly replaced by gypsum. During the salinification of separate parts of the sea in the Upper Cretaceous and Eocene of North Africa the dolomite stage was completely absent, but it was present in the development of the Fergana and Tadzhik basins. In modern sea lagoons the dolomite stage is completely absent, but in lakes having magnesium carbonate water, the precipitation of dolomite occurs. In this instance, too, precipitation occurs during the intermediate stage of salinification. The dolomite, after replacing calcite when the salinity was 3 to 4 percent, gives way to the separate deposition of magnesite and calcite when the salt content reaches high proportions [3], i.e., it repeats, as it were, in specific modern conditions that state of affairs which was widespread in the sea basins of the Paleozoic.

5. The formation of dolomites in antiquity and in the modern sea basins was distinguished not only by an increase in the salt content (apart from the normal salinity of the sea) but also by a pronounced increase in reserve alkaline and by a high content of CO_2 in the water.

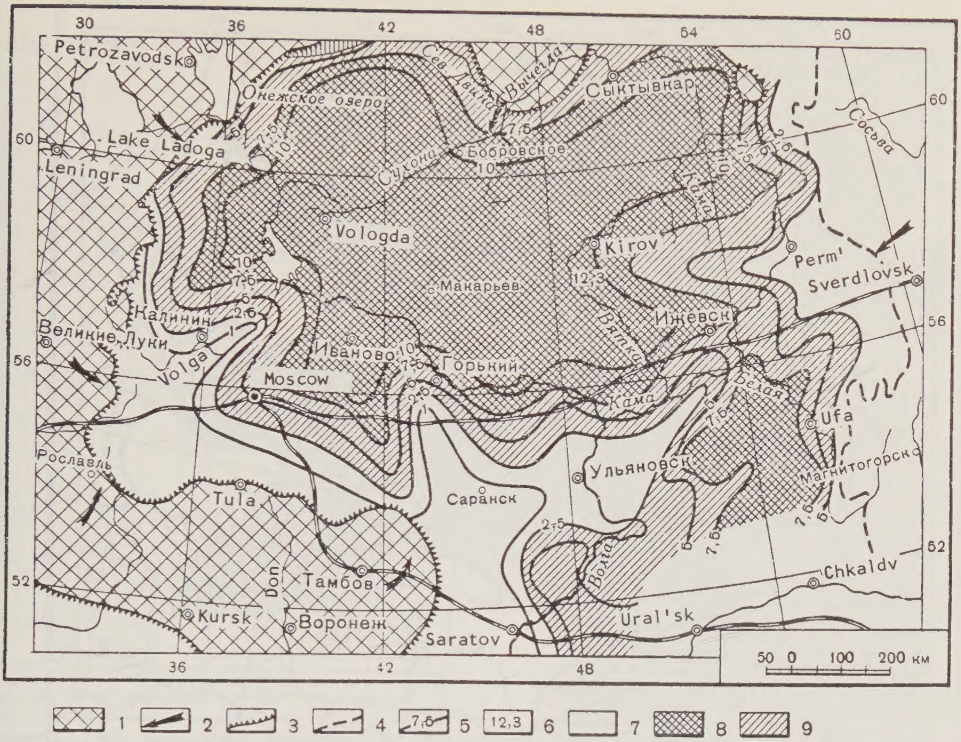

FIGURE 4. The Distribution of Magnesium in the Carboniferous Rocks of the Visean Stage of the Russian Platform. (according to A. B. Ronov, simplified)

1-Regions of Erosion; 2-Direction of the Transference of Fragmental material and the Probable Courses of the Discharge of Fresh Water; 3-Boundary of Regions of Erosion; 4-Boundary of Russian Platform and Urals Geosyncline; 5-Lines of Equal Mg Content (in %); 6-Average Mg Content in a Layer (in %); 7-Regions of Predominant Occurrence of Limestone; 8-Regions of Predominant Occurrence of Dolomite; 9-Regions of Occurrence of Carboniferous Rocks, Intermediate between Limestone and Dolomite.

These circumstances have been established beyond doubt not only by observations on modern dolomite-forming lakes, but also by data about the conditions under which dolomite has recently been successfully obtained experimentally. Thus, in the Balkhash lakes it has appeared in large quantities only when the alkaline reserve has reached 12 to 13 mg/equiv. to the liter, and pH = 8.8 - 8.9. This also applies in the case of the Burlinsky lakes and the soda lakes. These conditions are created by the concentration in the water of large amounts of $MgCO_3$ and also of Na_2CO_3 in the soda lakes [3].

Dolomite was obtained by A.V. Kazakov in ampules at a temperature of 150°C, in a solution containing 28 mgs CaO, 86 mgs MgO and 412 mgs of CO_2 per liter. In this case the reserve alkaline was 18.7 mg/liter; pH -6.66, and the relation of CO_2 to the final alkaline reserve = 3.07 [2]. It is understandable that under conditions of ordinary temperature the amount of reserve alkaline must be even higher. Recently, G.V. Chillingar gave a report on an interesting experiment for obtaining dolomite [7]: One liter of sea water was placed in a container fitted with an electric mixer. After adding 2.13g of $MgCO_3$ and 1.97g $CaCO_3$, (corresponding to the maximum solubility of $MgCO_3$ and $CaCO_3$ in sea water under a pressure of CO_2 under a pressure of 4 atmospheres) the container was placed in an autoclave with a pressure of 4 atmospheres of CO_2. The relation Ca:Mg in the residue formed upwards the end of the second week was 1.8. Although the particles which had precipitated were too small to determine their optical properties, the index of refraction proved to be that of dolomite (1.7). X-ray analysis also established the presence of

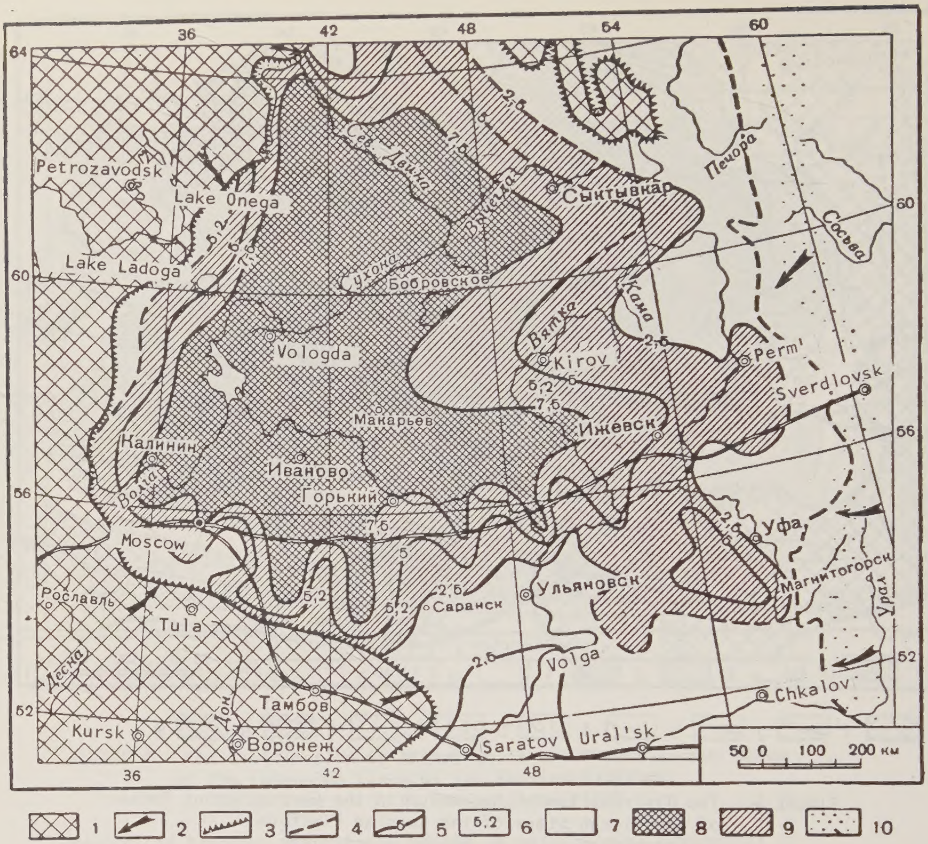

FIGURE 5. The Distribution of Magnesium in the Carboniferous Rocks of the Middle Carbon of the Russian Platform. (according to A. B. Ronov, simplified)

1-Regions of Erosion; 2-Direction of the Transference of Fragmental Material and the Probable Courses of the Discharge of Fresh Water; 3-Boundary of Regions of Erosion; 4-Boundary of Russian Platform and Urals Geosyncline; 5-Lines of Equal Mg Content (in %); 6-Average Mg Content in a Layer (in %); 7-Regions of Predominant Occurence of Limestone; 8-Regions of Predominant Occurence of Dolomite; 9-Regions of Occurrence of Carboniferous Rocks, Intermediate between Limestone and Dolomite; 10-Littoral Marine Terrigenous Deposits.

dolomite ([9], pp. 2261-2268).

As we see, both examples of this authentic synthesis of dolomite by an experimental process were carried out under conditions very far removed from nature. However, they both indicate, without question, the sharply increased significance of CO_2, and consequently of the reserve alkaline, for the precipitation of dolomite as compared to calcite. "The foregoing experiments of geologists and chemists in the precipitation of dolomite from sea water," writes G.V. Chillingar, "were fruitless because they did not apply high pressures to the CO_2."

Thus, on the basis both of experiments and natural observation it must be considere firmly established that the formation of dolo mite rocks took place (and still does in som places) as a result of the union of a rather high salt content with a sharply increased reserve alkaline and CO_2. From this it is clear that the role of salinity in any given instance is undoubtedly secondary; it only applies to the processes of the condensation of water which increase the reserve alkaline to the high degree necessary for the achieve ment of the saturation point for dolomite in a given solution.

6. Within the past six years it has become unmistakably clear that dolomite has evolved as the result of an irreversible

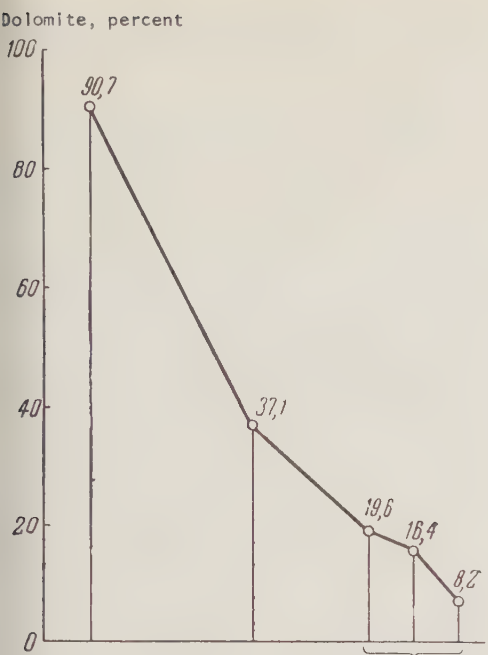

FIGURE 6. Mean dolomite content in rock complexes of different facies of the Lower Cambrian. (according to K. K. Zelenov)

rocess during the history of the Earth. It onsisted on the one hand of the "extinction" f a series of dolomite-bearing sedimentary icies, particularly such marine facies in ie arid zones and with lagoon, marginal iarine, pelagic marine formations so oundant in the Paleozoic era. On the other and, it took the form of a decrease in the itensity of formation of dolomite. The first ondition is readily visible on comparison of ie upper and lower tables on the diagram Fig. 1).

The first facts concerning the decrease in itensity of formation of dolomite in the arth's history were established by R. Daly y comparing organic samples of Paleozoic id Meso-Cenozoic rocks in the U.S.A. hat process was subsequently established ith great precision by A.P. Vinogradov, .B. Ronov and V.M. Ratynskiy [1] on ialogous rocks of the Russian Platform here a huge amount of material was exam-ed (Fig. 8a and 8b). D. Chillingar demon-rated it again in 1956 in another study of merican rocks [9], and similar data are available for the rocks of the Chinese Platform.

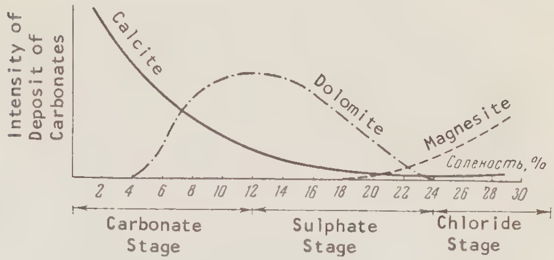

FIGURE 7. Variations in the Paragenesis of Minerals in the Process of the Salinification of the Kungursk Pre-Urals Lagoon.

The salinity of the water is represented on the abscissa; the intensity of calcite, dolmite, and magnesite deposition is shown on the ordinate. (Strakhov and Tsvetkov)

In 1956, when I was assessing the significance of these discoveries, I expressed the idea that the sharp decrease of dolomite in the Mesozoic and Cenozoic eras was a consequence of the fact that all three platforms -- North American, Russian and Chinese -- changed from arid zones of the Paleozoic to humid zones in the Mesozoic and Cenozoic eras. To confirm this it was necessary to collect data on the Carboniferous rocks Cr and Pg of North Africa where, at that time, an arid zone was located. This investigation was carried out by me in general terms only recently. Although I did not succeed in collecting numerical data, nonetheless it is clear and unmistakable from the ordinary geologic reports that there was no noticeable formation of dolomite in the seas and gulfs of the Mesozoic-Cenozoic arid regions. The formation of pure limestones, associated with gypsum took place there. This means that the sharp decrease in dolomite formation, referred to above by the American and Russian experts, truly reflects a process of the "extinction" of primary dolomite rocks in the past and their replacement by calcareous rocks. Doubtless this process had its origin in even more remote eras; only at the end of the Paleozoic did it become accentuated and noticeable.

These then, are the six cardinal, fundamental facts which must be considered by any modern hypothesis concerning the formation of dolomite. We shall see now to what degree the various modern theories fulfill this requirement.

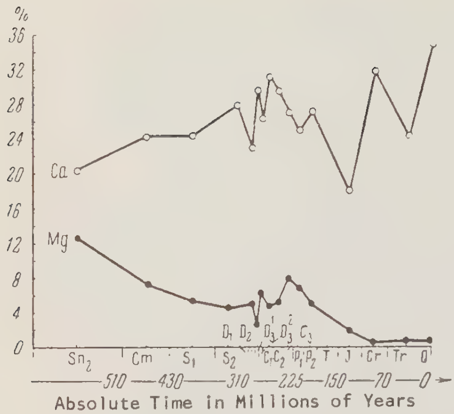

FIGURE 8a. The Variation with Time of the Average Percentage Content of Ca and Mg in the Carbonate Rocks of the Russian Platform.

(From the data of 7600 published analyses, and 198 analyses of established mean tests out of 8847 samples.)

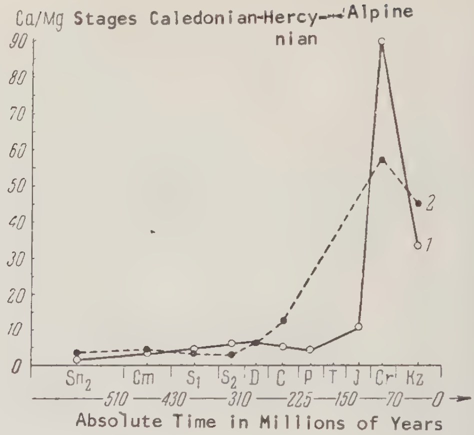

FIGURE 8b. The Curves of the Variation with Time of the Proportions Ca: Mg in the Carbonate rocks of the Russian Platform and North America.

1-Average Data from 8055 published analyses and 198 analyses of established average tests out of 8847 samples of Carboniferous Rocks of the Russian Platform; 2-Average Data of 853 samples of Carboniferous Rocks from North America (R. Daly).

II. Analysis of the Basic Hypotheses Concerning Dolomite Formation

At the present time three fundamentally different theories concerning the genesis of dolomite have been proposed.

According to the first theory (M.E. Noinskiy and S.G. Vishnyakov) [6], dolomites fall into two clearly defined genetic groups: bedded and spotted. The former are indicated by sedimentary formations which were deposited directly from water. The spotted dolomites on the other hand, are attributable to more recent, epigenetic processes which arose from the action of groundwater on limestone.

According to another theory, that of G.I. Teodorovich which is much more widely supported, the bedded dolomites are also regarded as primary and sedimentary formations while the spotted metasomatics which are far more prevalent in nature are regarded as diagenetic. Their formation in an early diagenetic stage boils down to the replacement of a portion of the Ca^{++} from the primary calcareous sediments by an ion of Mg^{++} which by one means or another entered from the bottom water. In this process, important roles are played by the Heidinger reaction; by the approximation to solution of $CaCO_3$ and $MgCO_3$ with the development of CO in the precipitate; and finally to the influence of the warm bottom currents.

The third theory was formulated in 1951-52 by me, after a study of the Carboniferous rocks C_3 of Samarskiy Luki [6]. According to this hypothesis, there are no genetic differences in principle between bedded and spotted metasomatic dolomites. Furthermore, in this and other cases dolomite in one way or another on the bottom during a stage of sedimentogenesis.

However, in the formation of bedded dolomites the admixture of $CaCO_3$ was ab-absent, or was present to only a negligible degree. On the other hand, in the development of the spotted metasomatics it was quite noticeable and sometimes even very great, whereupon both dolomite material and $CaCO_3$ were dispersed, at first very evenly. In the process of diagenesis, the dolomite component was energetically redistributed, leaving certain places and being concentrated in others forming lenses, spots, bosses, and other fantastic shapes. At the same time, in areas of secondary concentration, dolomite replaced calcite, having created metasomatic dolomite layers corresponding to the primary calcareous deposit.

The existence of so many explanations which differ so fundamentally compels one to

nalyze carefully how far each of them corresponds to the factual knowledge of dolomite rocks outlined in the preceding paragraph.

The hypothesis concerning the epigenetic development of dolomite rocks has three fundamental defects. If these dolomites had been truly epigenetic in origin, they could not have shown traces of their formation in a specific geomorphic situation. Nevertheless, such traces do exist. Metasomatic rocks in their predominant occurrences were indigenous to the basins of the arid zone, and even here only to those basins which showed traces of salinification. In this connection, it is impossible to understand why epigenetic dolomite was so widely developed in Paleozoic rocks, and so sharply curtailed in the Mesozoic (e. g., in the Cretaceous). Both cardinal facts of the history of the development of dolomite formations which have been mentioned, are not explained by the theory of epigenesis; they are simply ignored. Lastly, the epigenetic concept does not explain satisfactorily the purely quantitative side of the process. It must be borne in mind that the average dolomitization of the layers containing spotted dolomite is commonly very high, e. g., up to 40 percent in the C_3 Samarskiye Luki deposits. This demands that the epigenesis should yield huge quantities of magnesiu. What is the source of this?

The epigenetic hypothesis does not provide a satisfactory answer to this problem. In such situations that may be approached quantitatively, as for example, Samarskiye Luki [6], it appears that there is no source in nature. There is no place from which the amounts of magnesium necessary for an epigenetic displacement can be derived. The huge growths of spotted dolomite in nature cannot be explained by epigenesis. Only small specks, streaks and porous fillings in Carboniferous rocks could originate epigenetically, and then only in negligible quantities. In this way the choice must presently lie between the second, diagenetic, and the third, sedimentary hypotheses.

According to the theory of diagenesis, calcareous muds were precipitated in the first place, on the sea bottom, while the ingress of Mg into the precipitate took place during diagenesis by derivation of it from the bottom water. This interpretation, however, does not give a reasonable explanation either of the tendency of dolomite to form in the arid zones, of its general evolution in the history of the Earth, or of the interchange of the carbonaceous paragenesis during the salinification of the lagoon.

G. I. Teodorovich attempts to explain the attraction towards the salinified basins by the fact that on the salinification of sea water, some sort of approximation to solution took place between the $CaCO_3$ and $MgCO_3$, which was also propitious to the deposition of dolomite. However, as I pointed out recently [4], such an assertion is based on a misconception, and is refuted by existing experimental material. Nor does a tendency towards the Heidinger reaction help in this particular instance. As the recent experiments of M. G. Valyashko and his colleagues [4] have demonstrated, this reaction takes place only on the condition that $CaSO_4$ saturates the water. In sea water, however, even when the salt content rises to 5, 6, or even 7 percent, $CaSO_4$ is known to be far from the saturation point (it approaches it at a normal salinity of 15 percent). This applies even more so in the case of fresh water, depending on whether the sulphate ion in it disappears to a greater or lesser degree (or sometimes even completely) due to the action of desulphurizers. In the same way, the Heidinger reaction is eliminated as a possible source of diagenetic formation of dolomite. The tendency of metasomatic dolomite to form in the arid zones and during the stage of weak salinification of sea water (up to the point of its saturation with $CaSO_4$ and the precipitation of gypsum) remains unexplained in G. I. Teodorovich's approach to the problem. Teodorovich as yet has not tried to interpret the evolution of dolomite in the Earth's history, in the light of his hypothesis. This is not exactly accidental because the fact is that from his point of view these phenomena are incapable of explanation.

This theory has other weak points, too. It assumes a very energetic influx of Mg from the bottom water, so significant that the calciferous sediments are deprived of their Ca^{++}, by 1/2 or 3/4 (or even higher) and replaced by magnesium. The study of the diagenesis of modern sediments shows that, although the metabolism between the deposits and sea water actually takes place [5], it is manifested in very small quantities and cannot provide an explanation for such a sharply defined metasomatism as takes place in the development of spotted dolomite.

In actual fact, there exists only two ways in which magnesium may be taken up by the sediment from the bottom water. Firstly, it is a possible consequence of the reduction of the sulphate ion with which magnesium is associated in sea water. Let us assume that we have a marine sediment with a moisture content of 75 percent. On the full reduction of the sulphates 0. 26 percent sulphur will be retained in it, free to form a compound with iron pyrite; as a result, the pyrite content of the completely dry silt will be approximately 0. 5 percent. The normal pyrite content of argillaceous marine

sediments ranges from minute quantities to 1 percent -- rarely higher. In this way, on the formation of pyrite, there normally passes that ion of sulphate which is secreted with sea water in the mud, and only rarely are those portions of SO_4, which approximate in quantity the sulphate ion originally secreted, drawn into the mud by means of diffusion from the bottom water. This gives an idea of the smallness of the diffused transference of the sulphate ion from bed water to mud.

In addition to the sulphate ion, the magnesium and calcium associated with that ion, penetrate to the sediment from the sea water. It is possible to define by the following means, the quantities of magnesium involved. On the full reduction of the fresh-water sulphates, there was only 0.13 percent of Mg altogether in 0.26 percent of sulphur (on the calculation that only 2/3 of sulphur is associated with magnesium), which comprises 1/3 of the total amount of magnesium contained in fresh water at the moment of its secretion, and equal to 0.4 percent (in dry mud). Let us assume that all the magnesium was separated from the fresh water into the residue in the form of dolomite after the reduction of SO_4. This comprised a total general decrease in the concentration of magnesium in the mud solution by 1/3 in comparison with the bottom-water content, which, of course, cannot of itself cause an increase in the flow of magnesium ions from the bottom water to the mud; usually this decrease is still greater. On the other hand the precipitation of 0.13 percent of magnesium (from the weight of the mud) can generate an amount of diagenetic dolomite equal approximately to 1 percent of the weight of dry mud, i.e., a completely negligible quantity of a diagenetic dolomitization.

Another means of "pumping" magnesium out of the bottom water is the disintegration of organic matter by other groups of bacteria (except the desulphurizers) on the formation of CO_2.

As a result of this process, the reserve alkaline in the fresh water is sharply increased, but it falls again as the CO_2 leaves the residue. With the acute development of the reserve alkaline, the dolomite matter achieves saturation point and settles, in the process dolomitizing the carbonaceous sediment. This expulsion of magnesium from the mud solution could indeed be a cause of its inflow from the bottom water to the sediment. However, it must be borne in mind that the generation of diagenetic dolomite by the method described could not be achieved to any appreciable degree. Indeed, the disintegration of the organic matter which generates the CO_2 secreted in the silts only serves as a catalytic agent in the formation of dolomite. However, since the content of dolomite in silts, especially carbonaceous silt, is negligible, it follows that the generation by this means of diagenetic dolomite can only take place in quite minute proportions. All that could emerge from diagenesis caused by the method described would be isolated crystals of dolomite, small and rare dolomite spots, and nodules. In fact, however, the average dolomitization of the layers containing spotted dolomite commonly attained 30 to 70 percent, and even higher. The diagenetic processes could never have generated diagenetic dolomite on such a huge scale.

Teodorovich's theory (which was borrowed by him from A. Rivera) on the attraction of the warm bottom currents which enriched the sediment with magnesium, is of no further assistance in this matter. For such an enrichment it would surely be necessary to bring into action one of the concrete means of "pumping" magnesium into the silt I have just mentioned; we have seen that all these methods, separately or in conjunction, do not provide a solution to the problem. Furthermore, the shallow littoral zones of the seas are generally characterized by an increase in the temperature of the water. Therefore, according to Teodorovich's theory, one should expect an inevitable dolomitization, however great or small, of the littoral carbonaceous sediments, but such is not the case either in modern seas or in ancient ones. On the contrary, the maps of the dolomitization of the Carboniferous rocks which were drawn by A.B. Ronov, show the development of a high dolomite content occurring, not in littoral areas, but in the vast central part of the basins [6]. It is difficult to imagine these regions surrounded by warm underwater currents. It is quite evident that Teodorovich's theory concerning the dolomitizing role of warm currents is far-fetched, and does not correspond to the natural processes for the formation of dolomite.

Thus, a close inspection shows that although in theory the diagenetic formation of dolomite in sediments does indeed take place, in actual fact it does so only to a very limited degree, and cannot account for the genesis of spotted metasomatic dolomite in deposits which show any appreciable amounts of average dolomitization.

From all this it would appear that the only answer to the question of the genesis of spotted metasomatic dolomite rocks is the third concept outlined above; according to which there exists no fundamental difference in the formation of bedded or spotted dolomite. Both of these (and other types)

were obtained by the precipitation of magnesium-bearing substances directly from sea water. In the final analysis, the difference between the genesis of bedded and metasomatic dolomite lies in the incidentals of the processes, and not in their essence.

III. The Formation of Dolomite Rocks

The concrete mechanisms for the precipitation of magnesium compounds from water are regarded as having different in the Paleozoic and the very early eras in the history of the Earth.

As is known, the atmosphere in the Paleozoic was richer in CO_2, and as a result, the reserve alkaline of the sea was considerably higher than it is in modern times. Therefore, it may be granted, as I have often had occasion to point out, that in the Paleozoic, dolomite matter in the oceans was near to saturation in the water. Consequently, under the conditions in an arid zone, as soon as any section of the sea lost its free association with the main water mass, and became salinified to any appreciable extent, the dolomite matter quickly achieved the saturation point, and settled on the bottom in the form of a very fine-grained primary deposit. Saturation point was achieved all the more easily by the fact that the solubility of the dolomite was below the effects not only of $CaSO_4$, but also of $MgSO_4$ + $MgCl_2$ (three salts having analogous ions, which are present in sea-water), as a result of which the solubility of the dolomite, even in sea water of very low salt content, must have been diminished far more quickly than the solubility of $CaCO_3$. And if the diminution of solubility of $CaCO_3$ through the increased salinification of the water up to 1 percent was, as Trask maintains, 5 percent from the original, then in the case of dolomite, the corresponding figure should be at least twice, if not three times, higher [3].

Simultaneously with the deposition of dolomite the increase in salinity, brought about the complete extinction or sharp depression of pre-existing fauna and the appearance of a specific fauna. As a result, bedded dolomite emerged in the Paleozoic era having normal composition, and only became consolidated without undergoing any appreciable degree of trans-diffusion or metasomatism. These are sedimentary dolomites in the fullest sense of the term.

The genesis of spotted metasomatic dolomites was somewhat more complex. They emerged under a slightly lower degree of salinity than did the bedded dolomites. Consequently, the intensity of the primary deposit of dolomite matter from the water was appreciably lower in this case, which consequently resulted in a limited, and sometimes simply insignificant dolomite content in the fresh deposit. The dolomite in this instance was combined with a considerable quantity of biogenic or chemically-precipitated calcite, or simply formed an addition to it. Since in course of the very earliest diagenesis in the deposit, due to the action of bacteria, a great variety of physicochemical conditions emerged in relation to Eh, pH, and the concentration of the separate components of the muddy solution [5]; then in very remote times, under the influence of such conditions, there originated widespread migrations of different substances in the mud, which led to their departure from certain points and their concentration in others, all of which was accompanied by the displacement of certain substances from these points, or by a metasomatism of them. The dolomite which had originally been distributed in equal proportions in the deposit in the form of an admixture, also experienced a similar diagenetic transference in the sediments, as a result of which there emerged metasomatic concentrations of it in the form of lenses, specks, and spots in some parts of the deposit, and its complete disappearance in others.

Thus, the formation of metasomatic dolomite in the Paleozoic proceeded in two stages: viz., sedimentary and diagenetic. In the former, the dolomite was formed like a mineral precipitated from water of the basin; in the diagenetic stage, it assumed its present forms (spots, lenses, specks, etc.), and the lime dolomite rock acquired the petrographic appearance of spotted dolomite. Simplifying this complex, two-stage process in the formation of metasomatic dolomites of the Paleozoic, we propose to term them sedimentary-diagenetic.

The tendency of metasomatic spotted dolomites to form in seas of a somewhat lower salt content, in comparison to the bedded dolomites, determined parts of the sea in which they could form. In the peripheral, salinified zones of the seas they emerged chiefly in intermediate zones between the most mineralized water and areas of normal sea (Fig. 6). In the lagoons themselves, dolomite deposits occurred in littoral portions, along the periphery of the central, more salinified areas, if such existed, and also at the openings of lagoons, through which the currents were carried into less salty sea water. The most widely distributed dolomites of the Paleozoic were the metasomatic dolomites in the central, some what salinified parts of the Carboniferous seas of the Russian Platform.

The fact of the matter is that the insignificance and variability of the salinity of these parts of the sea made possible the deposition of dolomite, probably not every year, and even then only at certain times which corresponded to the warmest and driest periods.

One consequence of this was the emergence, as might be expected, of a blend of large and small quantities of $CaCO_3$ and dolomite. Under such conditions an immediate transformation of dolomite occurred, and metasomatism to $CaCO_3$, i.e., spotted, metasomatic dolomites finally emerged. Only in the strongly salinified lagoon sectors in the midst of a labyrinth of submarine shelves in the central regions of the Carboniferous seas, did salinification lead to the exclusion of $CaCO_3$ and to the deposition of dolomite throughout the year and for many years in succession. In these "lagoon basins" in the middle parts of the sea, normal, bedded, sedimentary dolomite then settled. In quantity, however, they were less than the spotted metasomatic dolomite, because the lagoons in which they had accumulated, occupied only very small parts of the pelagic, comparatively poorly salinified areas of the Carboniferous seas of the Russian Platform.

In this way, under conditions of high salinity, the intensive chemical deposition of dolomite from the water practically completely wiped out the deposit of calcite, and there emerged bedded dolomite without any traces of metasomatism to calcite, because there was nothing here on which it could have developed. With a decrease in the salt content of the basin, or parts of it, the primary deposition of dolomite from the water was diminished. Parallel to the deposition of dolomite, a precipitation of $CaCO_3$ went on at different seasons of the year; on the sea bed a blend of these minerals was formed, and a basis was created for the diagenetic transference of CaCO, and for the genesis of metasomatic, spotted dolomite.

The considerations outlined above go far to explain the formation of dolomite rocks through the initial salinification of Paleozoic waters of all types, but only partly apply to sea waters. But they still do not explain why at high salinity the dolomite deposit was replaced by a separate deposit of calcite and magnesite.

G.I. Teodorovich [7] attributes the change from dolomite paragenesis to that of magnesium and calcite as a result of the instability of dolomite in high concentrations of NaCl and magnesium sulphate. But what exactly are these concrete situations or reactions which render dolomite unstable in high salinity? This remains unexplained for the very good reason that the explanation itself in the past has simply been a formal mass of verbiage, in reality explaining nothing to the reader.

In interpretation of the change from carbonate paragenesis we must proceed from two facts: (1) the high solubility of dolomite compared to calcite and (2) the influences on a dolomite deposit of a simultaneous precipitation of gypsum.

In recent times a theory has been repeatedly propounded in the literature, according to which dolomite is less soluble than calcite (G.I. Teodorovich, O.K. Yanat'yev, B.S. Sokolov). However, the invariable emergence of dolomite after calcite in the sequence of salinification of all hydrochemical types of basins completely refutes this view, and testifies in favor of the high solubility of dolomite compared to calcite. Ignoring this fact, let us see how the origin of a gypsum deposit must reflect on the conditions of equilibrium affecting carbonates.

Since $CaSO_4$, to a much lesser degree than the carbonates, tends to give supersaturated solutions, and the supplies of the ion SO_4 in lagoon waters are vast, it is natural that from the moment of saturation of the brine in $CaSO_4 \cdot 2H_2O$ the precipitation of calcium from the brine begins to proceed in the direction of the formation of gypsum. This condition inevitably leads to the state where there is retained in the brine only such a quantity of Ca^{++}, as corresponds to the conditions for the saturation of the brine with $CaSO_4$, $2H_2O$, and $CaCO_3$; any excess over and above this amount cannot exist for long in the brine since it will be drawn off either by the deposition of the new portion of gypsum, or by the formation of crystals of $CaCO_3$, or by both of these processes together.

The saturation of water with dolomite to a higher degree than with calcite immediately postulates a degree of solubility such that the concentration of Ca^{++} will be higher than that which is defined by the saturation of water with $CaCO_3 + CaSO_4 \cdot 2H_2O$. The physical impossibility of obtaining such a concentration from the moment of origin of a deposit of gypsum is also the reason for the change over from dolomite paragenesis to that of magnesium calcite.[1]

[1] The importance of the presence in the solution of a sufficient quantity of Ca^{++} has been clearly demonstrated by A. V. Kazakov, who has pointed out that for dolomite, as distinct from magnesite, "the concentration of the (continued on following page)

For the formation of calcite it is necessary that appropriate amounts of $MgCO_3$ should be concentrated in the brine, i. e., its reserve alkaline increased from original marine conditions. This condition is fulfilled by Heidinger's reaction which, as has now been accurately established by M. G. Valyashko and his colleague (1950-52), begins to take place only after the $CaSO_4 \cdot 2H_2O$ has impregnated the water: [2]

$$CaCO_3 + MgSO_4 \longrightarrow MgCO_3 + \underline{CaSO_4 \cdot 2H_2O} \downarrow$$

Under the influence of this reaction the reserve alkaline is sharply reconstituted. $MgCO_3$ becomes its chief bearer with some minute portions of $CaCO_3$ as defined by the law of the dominant salt. Material conditions are created for the change from the deposition of gypsum of the dolomite paragenetic sequence to magnesite and calcite.

But if the considerations we have outlined are correct, we should discover in the gypsum and anhydrite, not dolomite, but a blend of dolomite with calcite and magnesite. Actually, dolomite is in fact present for the most part in anhydrite, alone or with a large quantity of dissolved calcite. In only a few instances do we find a blend of dolomite with magnesite and calcite. The complete "disintegration" of dolomite to $CaCO_3$ and $MgCO_3$ is recorded in rocks only much later, in cases of high salinity where NaCl and the potassium salts have already crystallized. This is the reason for the fact that the change from dolomite to a separate deposit of $CaCO_3$ and $MgCO_3$ is accomplished, not instantaneously as one might expect, but gradually. It may be asked how one can explain this "tardiness" in the change from the paragenesis of carbonate minerals in rocks and the gradualness of this change instead of a violent reaction.

As I see it, the reason for this should be sought for, not in the carbonate equilibria as such, but in the particular hydrologic system peculiar to all basins containing brine with a high mineral content.

The essence of this system lies in the fact that the concentration of salts does not remain constant throughout the year, but changes from time to time. In the autumn, and especially in spring, under the influence of snow and rain water, the brine is diluted to such a degree that the chemical deposition of salts ceases altogether; on the other hand, the deposited salts becomes soluble early in the year. In summer, during the period of maximum evaporation, the brine is heavily concentrated and the deposition of salts is renewed. Examining these yearly cycles in order to consider carbonate formations, it must be admitted that even if the basin also deposits $CaSO_4$ in summer, and has a corresponding carbonate regime, then in autumn, winter and spring, during the periods of dilution, the deposition of gypsum ceases, as we have already learned. In the first moments of the summer concentration of such diluted brine, up to the commencement of the deposition of gypsum, a precipitation of dolomite may occur. At a later period, when the gypsum has settled, the primary precipitation of dolomite ceases, and is succeeded by separate deposits of $CaCO_3$ and $MgCO_3$. On the balance of the yearly precipitation, a seam of small thickness is obtained in which are admixed in varying proportions the dolomite of the spring precipitation with the calcite and magnesite of the summer.

Just at the point in the yearly cycle when the lagoon is entering the sulphate stage, the spring dilution of its brine is very considerable and the period of its summer concentration up to the point where gypsum is formed, is extended for a long period, during which there will be a lot of dolomite in the deposit and very little summer calcite or magnesite. The amounts of the latter may be so small that neither a thermogram nor optical means can detect them. In actual fact, the beginning of the sulphate stage in the lagoon during the yearly cycle does not cause such a change in carbonate paragenesis as is postulated by the physico-chemical equilibria of the sulphate stage. With the passage of time, and with the normal increase in the salinity of the lagoon, the periods of spring dilution grow proportionately shorter, as also does their corresponding deposition of dolomite, and the intervals of the deposit of $CaSO_4$, $CaCO_3$ and magnesite become correspondingly more prolonged. In the sulphate deposits, this is referred to as the gradual reduction of dolomite and the increase of calcite and magnesite. In instances of very high salinity which usually indicate that the lagoon has already entered the

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[1] (continued from previous page) calcium must be in an order of magnitude greater than 50 m/l of CaO" ([2], p. 52) where $t^o = 150^o$; under lower temperatures it is certainly many times greater. Thus, it has been established by Kazakov that the theory whereby the precipitation of gypsum quickly cuts off the concentration of Ca^{++} in the brine and is a factor suspending the formation of dolomite, is borne out by experiment.

[2] $CaCO_3$ is being continuously brought into marine basins by river waters.

summer stage of chloride sedimentation, the spring dilution may be indicated by the fact that the brine will still be saturated with $CaSO_4$. From this moment the dolomite disappears completely from the composition of the halide deposits, and a typical paragenesis of magnesite and calcite begins. This paragenesis, as we see, is fully accomplished in the deposits of slightly later vintage than those deposited when the physico-chemical evolution of the lagoons created a prerequisite for it in the form of a summer saturation of brine with $CaSO_4$ and the beginning of the gypsum deposition.

It is appreciated that, depending on local geographic peculiarities, the moment of change of the paragenesis in the different lagoons commences at different times, and also that the magnesite-calcite paragenesis is associated now with rocks in the transitional stage from gypsum to halite; now with halite itself; and at other times with potassium salts. Furthermore, the transition itself from the dolomite to the magnesite-calcite paragenesis becomes gradual, fluid, and not intermittent. Thus the intrusion of a unique hydrologic system into the progress of the chemical accumulation of deposits in lagoons, creates conditions which at first glance give rise to contradictions between what is postulated by the accepted physico-chemical scheme and what is observed in the lagoon sequence of beds.

It is not difficult to see that, the more calcium a lagoon in which there are gypsum layers receives fluvial water containing bicarbonate, the more will be the Heidinger reaction, and the greater the quantity of magnesite which will accumulate in the gypsum rocks; some in the gypsum impregnated marls; others in rocks salinified with NaCl. At the same time the rocks are enriched with argillaceous material carried in by the rivers. The magnesite-bearing rocks in the Kuybyshev Zavolzh'ye region, which were recently analyzed and described by E. Frolova [8], are an example of this type of development.

All that has been written so far serves to explain the origin and the cessation of dolomite deposition by the salinification of the basins of the Paleozoic. Nonetheless, it is obvious that all these considerations are not applicable to those admittedly rare instances where formation of dolomite rocks occurred in the seas of the humid zones, for the most part in the areas of reef-building. It is obvious in this particular case that the role of the dolomite-depositing agent belongs not to salinification as such, but to other physico-chemical properties of the medium. The precipitation of dolomite did not occur either in those regions where the water was warmed to a marked degree, with a consequent lowering of the solubility of the dolomite, and its transformation to the stage of a saturated solution; nor in those areas of intense photosynthesis of phytoplankton or phytobentonite, which, in raising the pH, also brings about the saturation of the dolomite solution.

Both these causes of the precipitation of dolomite from water were certainly active in the salinified basins. But there they only incidentally hastened the precipitation of dolomite, which even without them went on as a result of the increased salinification of the water, i.e., they were a factor of occasional significance, but in the humid zones that they played a vital role.

It is understandable that the temporary and purely local increases of temperature and pH in sea water could only cause a smaller amount of dolomitization of the emerging deposit, because only metasomatic spots and lenses were formed here, and even these were neither pronounced nor numerous. Only where arid conditions occurred in certain parts of the platform basins and brought about a gradual salinification and a change in the salt content anomalous to that of the sea, did the processes of primary dolomitization become sufficiently pronounced to cause an increase in the size and numbers of the dolomite lenses. The classic example of this process is to be found in the Carboniferous of Samarskiye Luki.

And so, the bedded dolomite in the Paleozoic deposits represents primary formations. On the other hand, the spotted metasomatic dolomites are sedimentary diagenetic bodies, and those dolomites which filled the crevices, pores and caverns of fabricated rocks are epigenetic minerals.

In the Post-Paleozoic time vital changes in the processes of dolomite formation took place.

Since the content of CO_2 in the atmosphere sharply decreased as a result of the intensive accumulation of carbon and carbonates in the Upper Paleozoic it must be assumed that the dolomite matter in the sea water of the Mesozoic departed more from saturation than was the case in the Paleozoic era. The accumulation of carbonates in Jurassic, Cretaceous and Paleogene time and the carbonate deposits in Laramie time, could only intensify this very process.

It is understandable that in the Mesozoic and Cenozoic, precipitation of dolomite from its ever-diminishing saturated solution became more and more difficult, both in the

lagoons (because of their salinity) and in the sea by reason of an increase in pH and temperature. As a result, the formation of dolomite in lagoons and seas sharply declined -- even in the arid zones.

In spite of this, there still existed conditions, under which, and especially in lagoons, the formation of dolomite could be maintained at a high level (this also occurred in the seas of the arid zones). This state of affairs was created in those lagoons and portions of the sea where rivers or underground waters were depositing large quantities of $MgCO_3$, or indeed of $MgCO_3$ and Na_2CO_3. The incoming carbonate of magnesium (and sodium) increased the reserve alkaline in both the salt and fresh-water lagoons as well as the pH in the water, and brought the dolomite to the stage of saturation, in this way assisting its chemical deposition.

If there had been no introduction of $MgCO_3$ into the lagoon the deposition of dolomite was retarded or was completely absent. The instances of the simultaneous growth of lagoons, with and without dolomite stages, shows exactly this situation. Thus, in the Upper Jurassic (Tithonian) dolomite occurs in considerable quantities amongst the red-colored strata of the Northern Caucasus, side by side with gypsum. With the salinification of the Tithonian lagoons, the dolomites of the territory of Tajikistan did not settle, and the limestones were immediately replaced by gypsum. The difference in the composition of the waters supplied in the Caucasus and Tajikistan must be regarded as the cause of this. However, in Paleogene time in Fergana and Tadzhikistan, dolomite again made its appearance in quantity in the lagoon deposits. From this A.I. Osipova succeeded in proving conclusively the entry of waters abounding in $MgCO_3$ into the lagoons [6].

In this way, from an inevitable stage (arising directly from the salinification of the lagoons) deposition of dolomite occurred occasionally in the Mesozoic and Cenozoic eras, being absent only where there was a limited impregnation of the lagoon with salts (an introduction of $MgCO_3$ and Na_2CO_3).

All the attempts to obtain dolomite by the evaporation of solutions which were in equilibrium with the atmosphere were failures; instead of dolomite there settled admixtures $CaCO_3$ and the basic salts of $MgCO_3$ of variable composition. How may one explain this failure to obtain dolomite experimentally by means of evaporation in circumstances so clearly corresponding to natural conditions? Analyzing this problem in 1951 [3], I came to the conclusion that "the explanation lies in the fact that the compression of CO_2 in solution was insufficient, being less than 0.004 atmospheres, a factor which also compelled $MgCO_3$, with its tendency to give hydrated basic salts, to settle exactly in this form.[1] As is well known, the compression of CO_2 in mud increases during the process of diagenesis, and this causes the complete neutralization of magnesium carbonate; and t the redistribution of matter and the formation of dolomite" . . . ". . . if these considerations are correct, dolomite, with the concentration of CO_2 in the atmosphere of recent time, must always be a diagenetic mineral, even where (e.g., the Balkhash lakes) it accumulated in quantities, and gives the impression of primary sedimentation" ([3], p. 45).

But the present low content of CO_2 in the atmosphere is probably typical, and not simply a peculiarity of present geologic time. Its origins must be attributed without doubt, to the more or less remote past, possibly to the early Cenozoic, or even to the late Mesozoic.

In the same way, it must be admitted that in geologic history there has taken place not only a general diminution in the formation of dolomite from ancient times up to the present, but there has also been observed a fundamental alteration in the mechanism of formation of dolomite as a mineral. Instead of the spontaneous precipitation of dolomite from water in the Proterozoic and Paleozoic eras (and possibly at certain stages of the Mesozoic) at some stage or other the distinct deposition of $CaCO_3$ and the sub salts of magnesium carbonate began, with the consequent formation of directly precipitated dolomite.

The arrival of this moment indicated the complete "extinction" of sedimentary dolomites, and the "survival" of only sedimentary-diagenetic dolomite. Under these circumstances during abundant precipitation of the sub salts of magnesium in those lagoons which had received large amounts of $MgCO_3$ from the rivers, there emerged bedded dolomite which was approximated normal in composition. With the diminished precipitation of the sub salts of magnesium, typical spotted metasomatic dolomites emerged in the lagoons and especially the seas during slight, but general dolomitization of the Carboniferous strata; these deposits in no way differed from the Paleozoic types of this group.

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[1] This property of $MgCO_3$, already established in 1929 by the research of Klein, is clearly demonstrated in the experimental analyses of A.V. Kazakov.

In summary, it may be stated that during the past 10 to 12 years, our knowledge of the genesis of dolomite rocks has made great advances both theoretically and practically.

At the present time we are familiar with certain basic and decisive factors which were characteristic of the formation of dolomite in preceding eras; its multi-faciality; its partiality to arid zones; its attachment to the early stages of the salinification of basins and its replacement by magnesite and calcite paragenesis during time of high salt content; its inevitable presence in water with a high alkaline reserve and high CO_2; and finally, the sharp growth of the dolomite process in the Pre-Paleozoic and Paleozoic eras, and its marked suppression in the Mesozoic and especially Cenozoic eras.

A source has now been uncovered which explains all these basic factors. It consists of the regime of CO_2 and the reserve alkaline connected with it at various stages in the Earth's history, and in different physico-geographical circumstances. The accuracy of such a genetic arrangement has been proved by the first successful experiments and attempts to obtain dolomite in laboratory conditions by A. V. Kazakov and G. V. Chillingar, and also by the fact that the latter's experiments, as the author himself points out, were directed towards the verification of the fundamental principle in this case, and has vindicated it.

We may now consider, as a result of all this, that we are propounding reliable basic theories of the formation of dolomite, which allow us to apply the general principles governing the formation of dolomite to any specific instance, no matter under what conditions they arise. Future research must be directed towards establishing the details and development of these principles, both by means of lithologic field work, and by experiments under laboratory conditions which will approximate natural ones as closely as possible.

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BRIEF OUTLINE OF THE PRECAMBRIAN AND LOWER PALEOZOIC OF THE SCOTTISH HIGHLANDS

by

Ye. V. Pavlovskiy

This article is the first part of the work "Peculiarities of the Caledonian Mountains of Scotland." In the second part, which will be published in a future number, the early stages of the geologic history of this country and the role of plutonic faults will be discussed.

* * * * *

The idea that a large part of the territory of the northern half of Scotland -- the region of the Grampian and Northern Highlands -- is located within the Caledonian folded zone appeared in geologic literature long ago. The Caledonian platform (Eria) occupies a narrow belt along the northwest coast of Scotland and the archipelago of the Inner and Outer Hebrides. The border between the platform and the folded zone of the Caledonian in northwest Scotland is the line of the Moine thrust, along which the platform mould from southeast to northwest. It is considered that the Moine thrust, first located and studied by B. Peach and J. Horne [29], emerged and developed in the lower Paleozoic, forming a sharply marked border between the folded zone, called Caledonian[1] by E. Suess, and its platform (Eria) in extreme northwest Scotland.

Personal observations on the geology of Scotland at the time of the extended excursion organized by the International Association for the Study of the Plutonic Zones of the Earth's Crust (2nd session) in September 1957,[2] and also conversations with English geologists and the study of the literature permit, in the opinion of the writer, a somewhat different evaluation of the role of the Caledonian stage in the geologic history of the Scottish Highlands, emphasizing the importance of the more ancient, Precambrian tectonic movements, and revealing a definite sequence of Caledonian structures as well as their singularity.

In the southeast of the Scottish mountains lie the Grampian Highlands (Fig. 1), built of folded beds including rocks of the Moinian and Dalradian series and to a lesser extent Cambrian and Ordovician rocks in places overlain by rocks of Devonian age. The Grampian Highlands are bounded by the southern Border fault that separates the highlands from "Midland Valley." This "valley" is a deep tectonic depression composed chiefly of rocks of middle and upper Paleozoic.

In the northwest the linear zone of another large fault -- Great Glen -- serves as the boundary of the Grampian Highlands. This major fault zone was described in detail recently by W.Q. Kennedy [27]. Between Great Glen and the Moine thrust lie the Northern Highlands, formed almost exclusively of rocks of the Moine series, covered in the northwest (around Caithness, coast of Moray Firth) by Devonian formations, and in the southwest (Island of Mull, western end of the Ardnamurchen peninsula) by Mesozoic deposits and Cenozoic rocks of magmatic origin.

In the area where rocks of the Moine series occur, small inliers of older rocks of the Lewisian type are also exposed. The

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[1] The Latin name for Scotland was Caledonia.

[2] The following made up the staff of the Soviet delegation to the 2nd session of the Association for the Study of the Plutonic Zones of the Earth's crust: corresponding member of the U.S.S.R. Academy of Sciences, L.V. Pustovalov (head of the delegation), academician of the Kaz.S.S.R. Academy of Sciences, R.A. Borukayev, and the author of this article. The secretary-translator of the delegation was N.V. Khabarin.

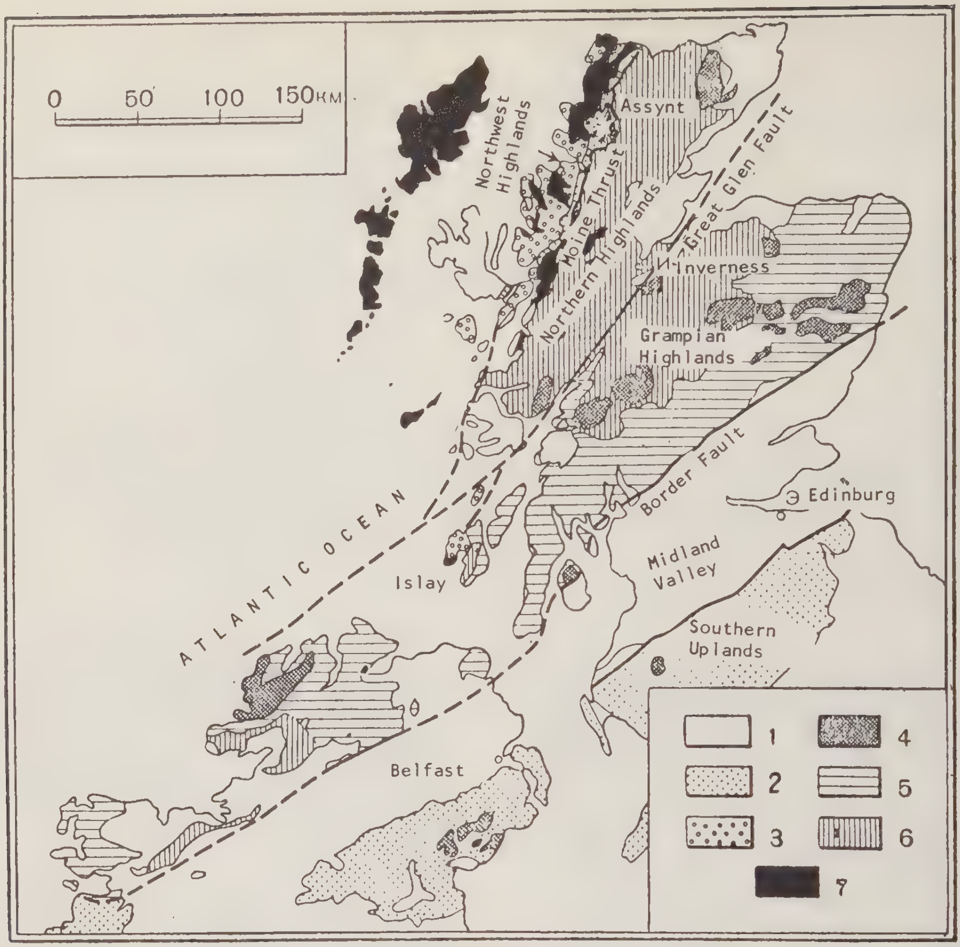

FIGURE 1. Geological sketch of Scotland and Northern Ireland after J.G.C. Anderson 10

1-middle and upper Paleozoic formations; 2-Lower Paleozoic; 3-Torridon; 4-largest intrusions (chiefly Caledonian); 5-Dalradian; 6-Moinian; 7-Lewisian.

foreland of Eria is composed of ancient Lewisian rocks covered with thick deposits of Torridon age. Cambrian and Ordovician rocks, which overlie Lewisian rocks in some places and rocks of Torridon age in other places, form a narrow belt, along the front of the Moine fault. The archipelago of the Inner Hebrides (including the Island of Skye) is made up of Tertiary lavas and to a lesser extent of sedimentary rocks of Mesozoic, Lower Paleozoic and Precambrian age.

As will be demonstrated, there are reasons for regarding Great Glen, the Moine thrust, and the Border fault as deep-seated plutonic faults, dislocations which as early as Precambrian and Early Paleozoic time broke up the territory of the Scottish mountains into a number of blocks extending in a northeast direction. From the northwest to the southeast of the ancient Eria platform these blocks are as follows: Northern Highlands, Grampian Highlands, and, outside the area of the present article, the "Midland Valley" block.

1. Eria Precambrian Platform

Two structural stages are well delineated. in the part of the Precambrian platform along the border of northwest Scotland. The lower stage -- the folded base -- is formed by sedimentary and igneous rocks of the Lewisian group. The complex and multiform upper

tructural stage consists of a sedimentary over of Torridon and lower Paleozoic age, nd also sedimentary and volcanic formations f Mesozoic and Cenozoic age [1, 2, 31].

A. Lower structural stage (base of Eria)

In a mixed complex of multiform crystalline schist and gneiss, of Lewisian age, two nain [31] groups of rocks are recognized -- rthogneiss and paragneiss (paraschist). In laces (Islands of Coll and Tiree), it has een established that orthogneiss cuts through ara-rocks. On this basis the paragneiss is onsidered to be the oldest part of the ewisian layer.

Besides the ortho- and paragneiss, partly heared dikes of basic and ultrabasic rocks f ancient pre-Torridon age are accompanied y very recent granite and pegmatite. heared granite, which constitutes a relaively young member of the Lewisian layer, ccording to J. Phemister [31], p. 12), in laces does not differ from oldest acid rthogneiss. From this it seems that the elative ages of ortho- and paragneiss are ot definitely known.

a) Paragneiss is known in the following egions of Scotland: Loch Maree, Gairloch nd Glenelg, and on the islands of Tiree, Coll, Iona (Inner Hebrides; [31]). The prinary sedimentary formations, components of nis layer of paragneiss, were clay beds commonly containing organic matter which vere later metamorphosed to granite), sandy nd carbonaceous rocks, and thin, ironearing layers. This series of marine deposts was transformed by metamorphism into ark, micaceous, garnetiferous and quartzitic chist, with seams of graphitic, crystalline mestone and dolomite. Stratified layers of ornblende and hornblende-chlorite schist lso occur in this same series.

In the Inner Hebrides, rocks that have merged from the metamorphism of arkose, andstone, dolomite and limestone are comnon. Other rocks of the Inner Hebrides nclude biotitic granulites and gneiss with arnet, multiform Ca-Mg- silicate rocks ontaining hypersthene as an accessory minral in places, and quartzitic-carbonitic ocks with scapolite. Feldspar and biotite of ne gneiss are superseded in places by rehnite. In the crystalline limestone, green yroxene and coccolite are present. Hyperthenic and eclogitic rocks of sedimentary rigin have been reported from the island of iree.

The para-rocks of the Outer Hebrides are ignificant in the present study. An alternation of thick and thin layers of psammitic, pelitic, and carbonitic rocks, believed to have been derived from sediments of shallow water origin, are present on Harris Island [26, 31]. These rocks may indicate flysch-like sedimentation. The series is composed alternately of quartzitic schist, multiform gneiss containing garnet, kyanite, cordierite and graphite, and diopsidic, forsteritic, paragatsitic, and dolomitic marble, in which graphite and phlogopite are also present. In Ca-silicate seams, scapolite is abundant. Garnet-bearing hornblende and pyroxenic gneiss and schist, considered by T. Jehu and R. Craig [26] to be metamorphosed basic lava and tuff, are interbedded with the multiform paragneiss. C. F. Davidson [15] considers these dark-colored rocks as metamorphosed sills of a gabbro-anorthosite complex.

It is significant that muscovite is not a characteristic mineral of the paragneiss of the Lewisian complex.

b) Orthogneiss is the most widespread type of rock of the Lewisian complex. It consists commonly of quartz, feldspar and one or more of the iron-magnesian minerals. In pyroxenic gneiss, the feldspar is oligoclase or andesine; besides monoclinic pyroxene, hornblende and biotite are present in lesser amount. There is feldspar in hornblende gneiss just as in the paragneiss described in the preceding section. Potash feldspar (microcline) abounds in the most acid biotitic and bi-micaceous gneiss. Bi-micaceous granulite [26], resembling the granulite of the Moine series of the Northern Highlands occurs in the Hebrides.

Isolated lenses and seams in the orthogneiss layers are of basic and ultrabasic composition. The ultrabasic varieties are biotitic-hornblende rocks, diopsidic rocks, containing pyroxene and hornblende, and rocks bearing spinel, olivine, and in places needles of anthophyllite; in this same group there is peridotite, serpentinite, dunite, and anthophyllitic-carbonaceous rocks. In the ultrabasic rocks garnet is common, hypersthene less widespread. In places the ultrabasic rocks have, in general, been metamorphosed into talc.

Rocks of basic composition form lenses, seams, or large masses in the midst of the orthogneiss. The main types of basic rocks contain pyroxenic-feldspar and hornblende-feldspar of different forms, usually massive or coarsely striated. Hypersthenic-augite and augite-feldspar bearing rocks with typical gabbro structure are known in this group. In hornblende-bearing rocks there is commonly garnet or epidote, and pale green pyroxene. According to C. Davidson [15] the basic and ultrabasic orthogneiss in the zone

of contact with younger acid orthogneiss on South Harris Island have undergone a marked change: the feldspar has turned to scapolite, the pyroxene has been replaced by hornblende, and the garnet has developed chloritic rims.

c) Later (but pre-Torridon) multiple intrusions are abundant in the area of occurrence of rocks of the Lewisian complex within the Eria platform. These intrusions cut the sheared ortho- and paragneiss. They may be divided into two groups of rocks: a) basic and ultrabasic and b) relatively younger acid rocks. The basic rocks (olivine norite, hyperite, gabbro, dolerite, diabase, enstatitic diabase, and epidorite) and the ultrabasic rocks (serpentinous pikrite, and dunite) usually form long dikes resistant to erosion. In places, these dikes are sheared; the basic rocks have turned into feldspar schist, and the ultrabasic into chloritic and talcous schist. Granite and pegmatite are very abundant. The form of the intrusions varies -- dikes, sills, and huge masses are present. The sheared granite of this group is shot through with unsheared varieties of the same rocks. The sheared granite is indistinguishable from the acid orthogneiss where the fracture surfaces of both groups of rocks are parallel.

The tectonics of the Lewisian complex is complicated and still insufficiently explained. The folding is most clearly defined. In the regions of Scourie and Torridon Bay, there are indications that a system of earlier sloping folded structures once existed. Numerous structurally complex zones and a change in strike have been noted [31, 38].

B. Upper structural stage of the Eria platform

a) The lower layer. A thick series of Torridon sediments forms the lower layer of the upper structural stage of the Eria platform. The chief area of development of Torridon deposits is situated in a belt 20-40 km in width along the southwest edge of the Eria platform. This strip extends in a northeast direction along the front of the Moine thrust. East of the thrust zone Torridon deposits are unknown.

A major nonconformity exists between the Torridon sedimentary rocks and the older Lewisian series. The complex, disjointed relief of an old hilly country buried under deposits of the Torridon series is clearly expressed (Fig. 2). The Torridon section for northwest Scotland, according to J. Phemister [31], is as follows:

In the southeast part of the Isle of Skye the cross-section of the lower Daiebeg formation is somewhat different. Here red rocks are not common; in the lower part of the sequence, green and yellow arkose and conglomerate, with seams of purple and green clay (60 to 90 m thick) occur in places; higher in the sequence, dark gray clay and arkose with lenses of calcareous rocks (180 to 360 m thick), overlapped by fine-grained arkose with seams of gray clay (750 m thick) are present. The formation is about 1,000 m in thickness where dark gray sandy clay and arkose with lenses of carbonaceous rocks occur.

Predominantly gray colors are also characteristic of à thick slice of the lower formation of Torridon age in the area bounded by the valleys of Loch Carron and Loch Alsh.

In other words, in the lower Torridon, red rocks disappear from the sequence in the zone of the Moine thrust, in the extreme southeast periphery of the Torridon depressions. On Iona Island, in the Torridon layer [31, p. 41] there are frequent alternations of sandy and clayey layers, i.e., there is an indication of a rhythm in the sedimentation.

Beyond the zone of the Moine thrust, feldspar-bearing sandstone and arkose of Torridon age have a breccia-like structure; in these beds, brecciated dark and light limestone remain unchanged. The dip is well defined. The clay, although compressed, have not acquired a real schistosity; in them, impressions of raindrops, shrinkage cracks, and ripple marks are well preserved. Traces of organisms are faintly defined; faint traces of worm crawlings are preserved [31] in the calcareous clay of the lower Daiebeg formation together with fine round phosphoritic nodules; brown phosphoritic filaments occur in the clay of the upper Aultbea formation.

In the lower and middle Torridon formation, seams bearing concentrated, heavy minerals (including magnetite) are present.

The composition of the clastic material of the breccia and conglomerate of the Daiebeg and Applecross formation is especially significant. The breccia consists of fragments of locally-occurring rocks of the Lewisian complex. The Applecross formation is present over a greater area than the lower Daiebeg formation and in places overlies rocks of that formation; in places it lies directly on rocks of the Lewisian complex. Exotic rocks, the original source of which is unknown, make up the Applecross conglomerate. Rock and mineral fragments within the Applecross conglomerate include quartz, quartzite, gravelly slate, jasper, arkose, felsite, and feldspar-porphyry.

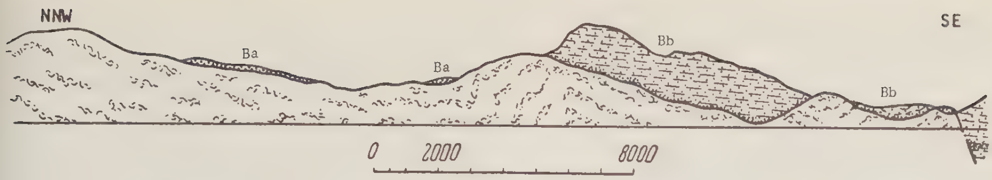

FIGURE 2. Nonconformity between Torridon sedimentary rocks and gneiss, after J. Phemister [31].

Ba--Daiebeg formation (Torridon,) Bb--Applecross formation (Torridon)

heared metamorphic rocks are present only arely as shingle. Spherulitic felsite, found 1 the composition of the shingle, in the pinion of J. Till [31], is identical with the elsite of the Uriconian series of Shropshire. licrocline and microperthite -- minerals which do not have wide distribution in the ewisian gneiss -- predominate in the composition of Torridon arkose. They were riginally concentrated chiefly north of the ria platform where ancient granite intruions are widely developed. Exotic rocks -- asper, aplite, red Na-granite and nordmarite -- have also been noted in the Torridon onglomerate that, together with local rocks f the Lewisian complex, makes up the hingle on Iona Island.

As mentioned previously, the thickness of the Torridon in various regions of Eria ranges widely. In the southern part of the belt-like area where Torridon rocks occur, this thickness is maximum -- about 6 km, but in the north at Cape Wrath, where the lower Daiebeg formation is lacking, the thickness of the series equals a total of 350-400 m. Sharp variations in the thickness were caused to some extent by irregularities of the substratum, but chiefly by the localization of the zone of greatest buckling in the southern part of the area of accumulation of Torridon sediments.

Sedimentary features of the Torridon, according to J. Phemister [31], are described

Name of formation	Thickness	Composition
Daiebeg	None in the north; at Gairloch -- 150 m, in the south of the Island of Skye -- 2100-2200 m.	Red sandstone, lower beds are dark argillite, gray sand, clay schist with lenses of carbonaceous rocks; breccia and conglomerate occur at the bottom of the sequence.
Applecross	In the region of Applecross -- 1800-2400 m, at Cape Wrath -- 300 m.	Chocolate and red arkose with seams of conglomerate; in places, chocolate and red clay, at the bottom -- breccia.
Aultbea	At Cape Coigach -- 900-1300 m; in the north, at Cape Wrath -- 75 m.	Alternation of sand, dark, sheared clay, and calcareous rocks, changing below to chocolate, red, and gray micaceous sandstone, with seams of gray and green clay.

On Lewis Island, in the depth of the Eria latform in the region of Stornoway harbor, ocks of probable Torridon age consist of onglomerate with lenses of chocolate sandtone. The total thickness of the series ere reaches 2700 m. The detritus and hingle of the conglomerate consist exclusively of local rocks of the Lewisian comlex which rocks of Torridon age overlie ith sharp angular unconformity. A slight ompression of sedimentary rocks occurred hroughout the Torridon series.

in the following manner. Extensive deposition of beds having an initial regional dip, concentration of heavy minerals in separate seams, ripples, shrinkage cracks, impressions of raindrops -- all these are believed to indicate a shallow-water regime with frequent drying of individual areas; strong currents may have occurred at times. These deposits accumulated in the environment of an alluvial lowland. The nonconformity between the upper formations and the older Torridon deposits of the basement in northern

Scotland, the orientation of the initial regional dip -- all these indicate that the dry land of this time lay to the north and northwest of the zone of accumulation of Torridon sediments. During deposition of the lower Daiebeg formation, part of northwest Scotland between Durness and Loch Maree was also dry land; subsequently it was gradually submerged during the course of the accumulation of sediments of the Applecross formation. On the dry land lying in the northwest, possibly a desert regime prevailed, evidenced by traces of aeolian processing of many shingles and also by the remarkable brilliance of the feldspar grains in sedimentary rocks of the Torridon series.

It should be kept in mind that the conglomerate of Port Askaig, developed on Islay Island and related earlier to the Torridon [4, 35], at the present time is considered part of the Dalradian series [10]. Therefore, there is no basis for associating this conglomerate of ice age origin with the time of the accumulation of Torridon sediments.

As stated previously, Torridon sedimentary rocks lie west of the Moine thrust zone, comparatively undisturbed. Up to the beginning of sedimentation during Cambrian time, Torridon deposits had a monoclinal dip of 10 to 20^{o} to the west [21, p. 38]. A clearly defined anticlinal structure having an almost meridional strike [see 31, Fig. 11, p. 40], developed in the coastal section of Scotland between Loch Maree and Loch Carron, west of the Moine thrust, in rocks of the middle member of the Applecross formation. Dislocated Torridon deposits occur in individual sheets of the Moine thrust. The huge syncline of Loch Alsh, tilted to the northwest at the side of the Eria platform, is another major structure which is traceable from Loch Alsh bay in the south to the region of the bays of Loch Carron and Loch Kishorn in the north. It involves gray-colored rocks of the lower part of the Daiebeg formation [see Fig. 3].

Torridon rocks are generally not metamorphosed. Only in the zone of the Moine thrust is the Torridon weakly metamorphosed where its sedimentary rocks were crumpled by folding and overturning at the time of Moine thrust movements. This is particularly true for the zone of the Loch Alsh syncline, for the regions of Torridon occurrence on the Island of Skye (Tarskavaig formation), and on the islands of Colonsay and Islay. Signs of metamorphism in these places include development of flow cleavage, conversion of clayey schist to phyllite and other phyllitic rocks, and the appearance of thin leaves of possibly recrystallized biotite [12, 35]. On the small islands southeast of Iona Island, Torridon clayey schist has been converted to chert on contact with Caledonian granite. In the region of Assint, Torridon rocks experienced contact metamorphism in the zones of contiguity with multiform post-Cambrian intrusions.

The Precambrian age of the Torridon is unquestioned because, as has long been known, Lower Cambrian rocks of northwest Scotland lie nonconformably upon rocks of Torridon age.

b) The upper layer. Rocks of the upper layer of the Eria platform are of Cambrian and Ordovician age. The distribution of Cambrian and Ordovician marine deposits in northwest Scotland is exceedingly unusual. They form a narrow belt 2 to 20 km wide, extending from about 160 km from Durness in the northwest to the southern end of the Island of Skye. This belt of lower Paleozoic deposits along the front of the Moine thrust is situated on the southeast edge of the Eria platform and extends parallel to the longitudinal axis of the Torridon depression.

Rocks of Cambrian age lie nonconformably on various Torridon stratigraphic horizons, or directly on the Lewisian gneiss. Conglomerate 0.3 to 3 m thick occur at the base of the sequence. The conglomerate includes

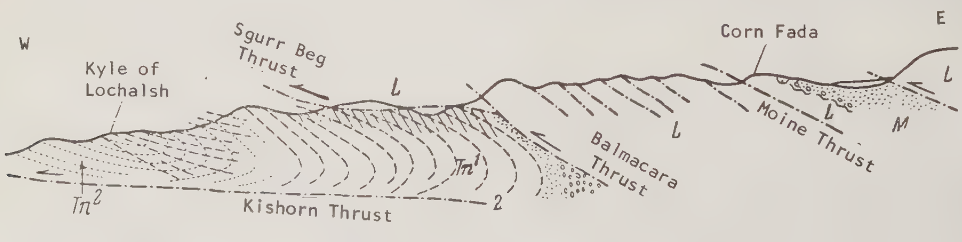

FIGURE 3. Section of the north coast of the bay of Lochalsh, after D. Kenanroy. Length of section 11 km.

L--Lewisian, M--Moine, Tn^1--Daiebeg formation (Torridon), Tn^2--Applecross formation (Torridon). Near the city of Kyle-of-Lochalsh, cleavage developed parallel to the axis of the surface synclines in the Applecross sandstone. Under the Lewisian cover (Sgurr Beg thrust), cleavage developed in sandstone of the Daiebeg formation as the result of shearing stresses.

ragments of quartz, feldspar, jasper, quartz-te, and felsite. Cross-bedded arkose and uartzite lie above the conglomerate and ogether with it form a layer of "basal uartzite" 45 to 100 m thick. The next layer of "pipe rock," mixed with massive and platy ine-grained quartzite, has a thickness of 75 o 90 m. This unit is remarkable for the bundance of vertical cylinders ("pipes") -- ossible traces of the life activity of worms; areful study of these forms makes it pos-sible to distinguish five zones within the unit. Higher in the sequence occur "fucoid layers" 12 to 15 m thick) -- carbonaceous clay, rgillite and arkose, on the bedding surfaces of which worms have left traces of their ctivity. Here also is contained the faunal omplex of the Lower Cambrian Olenellus one (tribolites, brachiopods, and mollusks). Arkose and quartzite with seams of carbona-eous clay -- "serpulite arkose" (9 m thick), with Olenellus and Saltarella, lie still higher n the sequence. The next younger strati-graphic unit is the Durness limestone and lolomite, 460 to 480 m thick. In the upper art of the Durness beds there are brachio-ods, tribolites, and sponges, identical to hose forms characteristic of the Beekman-own-Canadian rocks of North America which n the opinion of E. Ulrich and C. Shukhert 25], span the interval from Upper Cambrian o Lower Ordovician. O. Jones [25] considers t indisputable that the upper part of the car-onaceous unit belongs to the Ordovician. He uggests the possibility of the lack of Middle nd Upper Cambrian deposits in carbonaceous equence, similar to that observed in many ower Paleozoic sections in North America e.g., in Vermont and Massachusetts).

In carbonaceous rocks, well-rounded sand rains, possibly of aeolian origin, are common. n different parts of the sequence, stromatolite ones, siliceous concretions and lenses are ob-erved. J. Phemister [31] considers that the urness carbonaceous rocks were deposited s slime, ooze, and silt sediments that ccumulated so slowly that many hard parts f organisms were dissolved before they ere covered by additional new sediments.

In the opinion of C. Poulsen [32], the ambrian-Ordovician profile of northwest cotland displays complete stratigraphic con-ruity with the analogous deposits of Eastern reenland. O. Jones [25] emphasizes the emarkable similarity of the fauna of Scot-and of the lower Paleozoic to the contempor-neous fauna of the North American "basin" f the St. Lawrence, which includes New-oundland.

There were noteworthy environmental con-itions in the deposition of the lower Paleo-oic sedimentary series of northwest Scot-and. East and southeast of the zone of the Moine thrust the Cambrian-Ordovician rocks lie with a monoclinal dip of 5^{0} to 20^{0} of the surfaces. Within the thrust zone, in Allochton, the lower Paleozoic sedimentary rocks also maintain their monoclinal dip, in places. Small, sporadically distributed folds whose axial planes dip northwest, have developed in individual thrust sheets. These folds are open or firmly compresses, and, in places, they are overturned.

In the region of Assint, in the Moine thrust zone, just as in the underlying Torridon and Lewisian rocks, the Cambrian has been intruded by a series of sills, dikes, and laccoliths of various composition. The sills and dikes are composed of felsite, porphyry, and lamprophyre. All the intrusions resemble, in general, the Lower Devonian intrusions of central Scotland. One notable difference is that the rocks of Assint are richer in alkalies, a fact emphasized by the appearance of aegirite in the most acid varieties. Laccoliths, which occupy areas of a few square kilometers, are of mixed composition. They include quartzitic syenite bearing nepheline, sodalite, aegirite and riebeckite, and alkaline ultrabasic rocks (melanite-bearing pyroxenite) and shonkinite. The relatively later dikes and sills are of mixed alkaline syenite, alkaline granite, alkaline felsite, and cancrinitic rocks. In the zones of contact with alkaline laccoliths, the Cambrian carbonaceous rocks have turned to diopsidic, forsteritic, tremolitic, and brusitic marbles; in places -- in wollastronitic rocks and in skarns containing zoisite -- there are idocrase and grossularite, and also micas, tremolite and diopside.

* * * * *

Within the limits of the present work we cannot touch on the description of those geological events affecting the Eria platform during later Ordovician time. We shall describe the role of the Moine Caledonian thrust and some of its peculiarities in a later part of this paper.

II. The Northern Highlands structural block

The Northern Highlands block, bounded on the northwest by the Moine thrust and on the southeast by Great Glen fault, is a complicated tectonic structure. Two structural stages are discernible. The lower stage is formed by the folded Lewisian complexes, sub-Moine and Moine, and the upper stage by Devonian formations and rocks of the Mezo-Cenozoic.

A. Lower structural stage of the Northern Highlands block

The lower structural stage of the Northern Highlands block may be subdivided into a lower layer, the folded base (Lewisian and sub-Moine complexes), and an upper layer -- the folded Moine complex.

a) The lower subdivision is divided, within the Northern Highlands, into small areas, the so-called "windows" -- cells of anticlinal structure. These "windows" are comparatively small in area and are concentrated chiefly where the Northern Highlands are adjacent to the Moine thrust zone on the southeast; they also occur in the central part of the Highlands. Some of these "windows," formerly believed to be mixed rocks of the Lewisian complex (e.g., at Scarderoy and around Ross-shire), are now thought to be metamorphosed rocks of the Moine series and not ancient Lewisian gneiss at all (J. Sutton and J. Watson [39]).

In the vicinity of Ross and Inverness, biotitic and feldspathic gneiss, feldspathic, graphitic, disthenic, micaceous, and other schist, and cipolino and other marble, developed from rocks under the Moine series. Basic and ultrabasic rocks are present in abundance as pseudomorphic bodies (eclogite, serpentinite, and talc schist). A nonconformity is clearly present, in places, between the Moine and Lewisian complexes in the "windows," as, for example, near Glenelg on the coast of Sleat Sound and in the vicinity of Strome Ferry on the south coast of Loch Carron. Within these regions, thin-layered metamorphosed conglomerate with crushed pebbles consisting of rocks of the Lewisian complex are part of the basal Moine series. In many other "windows," conglomerate of the Lewisian complex may not be present at the base of the Moine series [34]. The conglomerate and Moine schist nonconformably overlay the underlying gneiss, as for example, in the region southwest of Glenelg [13].

A conspicuous conformity of Moine and Lewisian folded structures is generally apparent. This phenomenon is clearly marked on the detailed map of the region of Glenelg.

It is believed that the local folding here would not have been so intensive, if the folding had been more widespread.

A separation of the Moine from the Lewisian is difficult in places because the Moine schist is metamorphosed to various stages of gneiss. Foliation of the Moine gneiss and the Lewisian rocks is generally parallel. Under such circumstances, separation of the Moine and Lewisian complexes is practically impossible [31, p. 18], and consequently there is no basis for positive identification of the lower structural layer.

West of the Inverness district, a group of sub-Moine rocks is distinguished by J. Richey and W. Kennedy [36]. These are striated and massive gneiss of ultrabasic, basic, and acid composition, interstratified with thin-layered quartz-muscovite-epidotite granulite and biotite-epidote-garnet schist. The overlying Moine series rests nonconformably on various members of the sub-Moine sequence. Cross-bedding is present in the sub-Moine granulite; consequently, it is possible to construct a clear section of the series.

J. Phemister ([31], p. 20) prefers to consider the paraschist and orthogneiss of the Moray region as part of the Moine series, resembling the feldspathic gneiss and their associated granulite of the lower Moine series in the northern part of the Northern Highlands.

b) The upper structural subdivision is the Moine series in the Northern Highlands. Its composition is surprisingly uniform. Two types of rocks are characteristic of the Moine series -- quartzitic-feldspathic psammitic rocks, called granulite by English geologists, and pelitic rocks, usually represented by micaceous schist. J. Richey and W. Kennedy [36] separated the Moine series of the Moray region into: a) lower psammitic group, b) striated and pelitic group, and c) upper psammitic group. J. Anderson ([10], p. 12]) believes this profile, although not applicable to all areas of the Northern Highlands, is essentially correct, also for Ross-shire, where an "upper pelitic series" is located above the "upper psammitic series."

According to J. Phemister [31], calcareous siliceous granulite is subordinate to the predominating psammitic and pelitic rocks in the Moine series, and dolomite and limestone occur in very small quantity. The rocks of the psammitic type here include quartzite, gray quartzitic-feldspathic granulite, bearing biotite, muscovite, and garnet. Zoisite or epidote, apatite, zircon, sphene, and iron-ore minerals are the accessory minerals. The structure of the granulite is granoblastic, the texture schistose, more rarely massive. Subordinate to the granulite are very thin seams (as much as 1.5 cm thick) consisting entirely of round grains of heavy minerals -- zircon, garnet, sphene, epidote, allanite, ilmenite, and magnetite.

The basal conglomerate of the Moine series, rarely found, as mentioned above, consists of fragments of quartz, feldspar, and various types of gneiss. The cement of

the conglomerate is an epidote-biotite schist, in places bearing hornblende and garnet. In the region of Glenelg the conglomerate is exposed for 5 km along the strike. Its thickness is 6 to 9 m.

Other conglomerate units demanding attention are those associated with the middle part of the Moine series and with the "upper psammitic series," occuring in the central part of the Northern Highlands west of the granite mass, Carn Chuinneag. In the shingle of these conglomerates, besides quartz, alkaline feldspar, quartzitic-feldspathic granulite and, rarely, hematitic-micaceous schist, there is also quartzitic felsite. As J. Phemister emphasizes ([31], p. 17]), the granulite of the upper psammitic series is substantially less recrystallized than the average Moine granulite.

The pelitic type of rocks of the Moine series is chiefly micaceous schist with a sharply defined schistose structure. Where quartz and feldspar are present in these rocks in considerable quantity, the light minerals are separated in individual layers. The common schist minerals, garnet and, more rarely, kyanite and staurolite, are also present. In zones of massive granite injections, sillimanite is also found in the schist.

Semi-pelitic schist occurs as interstitial links between micaceous schist and quartzitic-feldspathic granulite. A subordinate role in the composition of the Moine series is played by zoisitic granulite bearing garnet and hornblende, epidosite bearing pyroxene or amphibole, calcite-garnet granulite, pure marble and multiform calcium-silicate rocks bearing green amphibole, diopside, zoisite, calcite, quartz, sphene, microcline, and andesine. At the contacts of marble and later granite intrusions, wollastonitic, scapolitic, and other rocks have developed.

The great thickness of the Moine series is apparently not uniform everywhere in the Northern Highlands. The thickness of the three lower series is 1200 m in the region of Fenig [Fenwick?] Forest; the thickness of the upper series is approximately the same ([31], p. 21). In the southwest, in the region of Arisaig and Morar [36], the thickness of the series is different; the lower psammitic beds are about 1000 m, the striated and pelitic beds about 1000 m, and the upper psammitic beds about 3600 m thick.

Primary clastic textures, cross-bedding, ripples, shrinkage cracks, and the tracks of marine organisms are all well preserved in the Moine series. Cross-bedding indicates that sediments were derived from the north-northeast. J. Flett believes that sediments constituting the Moine series rocks were originally red, and that the ferrous coloring substance went into the composition of biotite during the process of a subsequent metamorphism [31]. In his opinion, the whole series is of continental origin, the sedimentary environment during its deposition resembling that which existed during Torridon time or similar to the environment during accumulation of the Old Red sandstone of the Devonian system. J. Phemister opposes these views ([31], p. 20). In his opinion, the Moine series was formed under estuarine or lagoon conditions.

The tectonics of the Northern Highlands has still not been fully explained. It is clear that the regionally metamorphosed Moine series constitutes a complex system of tightly compressed folds. The morphology of the folded structures is also extremely complicated. An inverted stratification in many exposed sections indicates the great role of horizontal folding in the structure of the Northern Highlands. On the other hand, in the regions of Caithness and Sutherland, steep isoclinal folds predominate. The axial planes of these folds dip, in general, to the east or southeast. The isoclinal folds of a second cycle are superposed on steeper folded structures. The axes of these superposed folds are subparallel to the axes of the folds of the first cycle. The joints of the isoclinal folds commonly undulate. Sometimes in places, the isoclinal folds have a different direction of dip than the joints that are bent in the form of the letter S.

The regional strike of the folds in northern Argyllshire is northeastward. In the districts of Inverness and Ross, where "windows" of rocks of the Lewisian complex have developed, the north-northeast strike of the folds is replaced by faulted sigmoids whose occurrence is explained by the development of an overlying system of folds with a northwest strike. Near the front of the Moine thrust, the strike of the folds is determined by the strike of the surface thrust; in this area, a latitudinal strike is common. In the region of the Assynt "shelf" of the Moine thrust in the east, the strike of the folds of the Moine series trends to the southeast. Horizontal compression at a right angle to the direction of movement along the Moine thrust is suggested also by the rod structure of the schist (e.g., in the region of Ben Sutaig and other areas) east of the thrust front. These "rods" emerged because of the squeezing of shallow isoclinal folds trending parallel to the direction of movement of the Moine thrust ([31], p. 40). The great complexity of the folded structures of the Moine series is well demonstrated by R. Harker, who has recently studied the central part of the Northern Highlands [23].

Between the bays of Loch Alsh and Loch Hourn east of the front of the Moine thrust, three stages of folding have been distinguished by English geologists in Moine and Lewisian rocks. The older system of strongly compressed folds has a bow-shaped strike (north-northeast -- meridional -- northwest). On this system of synclinal and anticlinal folds are superposed the folded structures of the second and third stages (termed synform and antiform). Structures of the second stage are S-shaped. The axes of these folds strike from south to north: northeast, west-northwest, north-northeast. The axes of the overlying folds of the third stage trend chiefly in northeast, approximately parallel to the front of the Moine thrust. The three-stage development of folded structures east of the Moine thrust front in the region of Glenelg is clearly defined on sheet 1-m of the geological map of Scotland at the scale of 1 mile to 1 inch.

The writer suggests the possibility that the zone of migmatization (incomplete granitization), within which there are some ultrametamorphosed, generally concordant, granite bodies, is associated to some extent with the emergence at the surface of the most deep-seated parts of the Moine series. In other words, it is possible to view the zone of migmatization and granitization of the Northern Highlands mixed with rocks of the Moine series, generally speaking, as the core of a huge anticlinorium, extending in a northeast direction. If this assumption is correct, then, on the whole, the tectonics of the Northern Highlands, notwithstanding all its complexity in detail, can be understood as the structure of a huge anticlinorium in the core of which the processes of metamorphism and ultrametamorphism are expressed in the greatest degree. In both flanks of the anticlinorium, the intensity of metamorphism of the Moine schist decreases correspondingly.

A number of magmatic rocks are associated with the folded Moine complex of the Northern Highlands. The intrusions which arose before the development of schistosity (pre-foliation intrusions are relatively older). The blocks and stratified bodies of ultrabasic composition -- serpentinites, epinorites, in places converted to tremolites, and chloritic and talc schists can be dated from this time. The basic rocks of this group occur in the form of blocks, dikes, and stratified layers, and are represented by amphibolites, epidolerites, epidiorites, and appinites. The relatively older acid intrusions are quite rare; the most remarkable examples of these are represented by the masses of Carn Chuinneag and Inchbea, situated in the central part of the Northern Highlands. The composition of these masses is complex: the chief rock type here is the biotitic granitic augen gneiss. The "eyeglasses," which consist of crystals of orthoclase or microcline, extend in one direction and merge into deeply marked strips parallel to the schistosity of the surrounding Moine schist. The peak of Carn Chuinneag is made up of a strip of alkaline riebeckite-aegirite gneiss.

A wide belt of massive acid injections, extending from southwest to northeast [27, 28, 31] is associated with this same central part of the Northern Highlands. Within this belt the Moine schist is converted to migmatite owing to a massive, thin, sheet-like injection of granite and pegmatite. In places (Strait Halladale) in this zone there is porphyritic granitic gneiss, concordant with the surrounding gneiss. In the dikes there is granite, pegmatite, and aplite.

Younger, but pre-Middle Devonian intrusions are rather widespread in the Northern Highlands. The alkaline syenite intrusions of Loch Loyal are also contemporaneous, and are related to the post-Cambrian syenite of the Assynt region, the dikes of sheared lamprophyre, the small intrusions of appinite, and the blocks and masses of unsheared and undisturbed granite, granodiorite and diorite (largest of them -- Helmsdale, Lairg, Strontian, and Ross of Mull). The latter in their chief mass are composed of augitic diorite, granite and syenite, adamellite and tonalite.

The age of the Moine series is evaluated differently by various investigators. The Moine series has been correlated with the Lewisian complex [31]; other workers have considered its age as post-Lewisian but pro-Torridon; a third group believes it is Torridon, and a fourth post-Cambrian. From all that has been said previously about the Moine series, it follows that its correlation with the Lewisian complex is inadmissible at the present time. This series is certainly younger than the Lewisian complex. Its pre-Torridon age, as J. Horne believed in his day, also cannot be sufficiently supported at the present time. These opinions were connected with an incorrect assumption about the Precambrian age of the Dalradian series. In the light of the studies of the last decades, even the point of view that the Moine series is post-Cambrian has been dropped. Only G. Frodin [17] supports this view, but English geologists rejected it after the Moine thrust zone was studied and the relationships among the Lewisian, Moine, Torridon and Cambrian-Ordovician complexes in northwest Scotland were explained. At the present time, the view prevails that the Moine is a stratigraphic analogue of the Torridon. First expressed by B. Peach [29, 30], this viewpoint is supported by W. Kennedy [27, 28],

E. Bayley [12], C. Clough [14], J. Phemister [31], J. Anderson [9-11], O. Holtedahl [24], A.A. Polkanov [3, 4], and others. The lithologic difference between the Torridon and the Moine is explained by the varying conditions of sedimentation.

What is the age of the Moine series folds, of the regional metamorphism of this series, of the phenomena of massive injection and various intrusions? There are many differing answers in English geologic literature to this complex and difficult question, and these answers, naturally, are connected with definite ideas about the age of the Moine series. As described in the preceding pages, there is a basis for comparison of the Moine series with the Torridon. Consequently, all the ensuing phenomena -- folding, regional metamorphism, massive injection, and development of various intrusions -- appear to be post-Moine and post-Torridon. On the other hand, the upper age limit of this group of phenomena is determined by the fact that the relatively youngest intrusions of the Northern Highlands (Loch Loyal syenite, granite intrusions of Helmsdale, Lairg, Strontian, Ross of Mull, et al.) are pre-Middle Devonian in age [31]. In evaluating the age of the Moine series folding and all accompanying phenomena in the Northern Highlands, it is necessary to consider the essential fact, as emphasized by J. Anderson [10, 11], that in the more southern regions of Scotland, within the Grampian Highlands, the upper Moine is overlain by lower Dalradian, with no traces of discontinuity or nonconformity, and then by upper Dalradian, which is equivalent to Cambrian. We shall return to this important point again later after describing the ancient rocks of the Grampian Highlands.

E. Bayley [12], J. Phemister [31], J. Anderson [11], like many other British investigators, on the basis of careful analysis of the facts, consider that the Moine series folding, the regional (dynamo-thermal) metamorphism, the development of schistosity in the rocks of this series, their migmatization, and the development of early intrusions, are all associated with the initial stages of the Caledonian orogeny. The formation of the Moine thrust is referred by J. Phemister [31] to a later stage of the same Caledonian cycle.

The Moine thrust developed later than the accumulation of the Cambrian-Ordovician layer in northwest Scotland, after the introduction of various alkaline intrusions of the Assynt region. Only certain lamprophyres, related to the surface of the thrust, are younger than the Moine thrust. Various metamorphosed rocks, deformed in the process of development of the thrust, are found in the shingle of Middle Devonian conglomerate, by which the age of the Moine thrust is determined as pre-Middle Devonian ([31], p. 51; see Fig. 4).

B. Upper structural stage of the Northern Highlands block

The upper structural stage of the Northern Highlands block consists, as mentioned previously, of Devonian, Carboniferous, and Mezo-Cenozoic formations. The Northern Highlands Devonian system, represented by middle (Orcadian series) and upper divisions, rests nonconformably on the underlying rocks. The middle series is composed of arkose, conglomerate, clay, in places bituminous sandstone and thin seams of limestone with numerous fossils of fish, terrestrial plant-life (in places, Estheridae). Individual layers are red, chocolate, and grayish. The thickness of the Orcadian series is notable; in the district of Caithness it reaches 5500 to 5900 m. Rocks of the Moine series, Torridon and Cambrian, and also various igneous rocks of "ancient" and "young" intrusions which cut the above-mentioned sedimentary series are found in the shingle of the conglomerate.

The upper division of the Devonian is less extensive. It is composed of rose-colored, yellow, and red sandstone and arkose with fish fossils. Its thickness reaches 600 m.

In both Devonian series, ripples and shrinkage cracks are preserved on bedding surfaces. Cross-bedding is well defined in places. Structural weakness of individual components of the Devonian sequence have been noted where there are local disturbances. Sedimentation occurred in fresh water or saline basins. The Devonian deposits of the Orcadian syneclise are very flat-lying; in places they form sloping wave-like folds and are cut by later faults and lamprophyre dikes.

III. The Grampian Highlands Block

This block, the largest in Scotland, is bounded on the northwest by the linear fault of Great Glen and of the south by the line of the Border fault which separates the Highlands from the Midland Valley. In the block structure, just as in the preceding case, two structural stages are clearly delineated. The lower stage is composed primarily of a thick series of sedimentary rocks, partly of volcanic rocks of the Moine, Dalradian, and Ordovician series, together with various intrusions cutting through them; the upper (multi-layered) stage is composed of Devonian, Carboniferous, Permian, and Triassic rocks, cut by dikes of various composition,

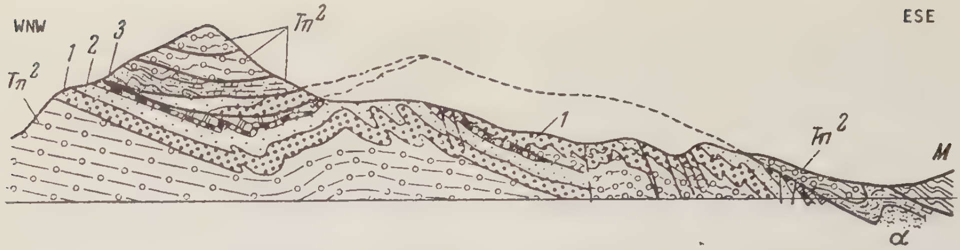

FIGURE 4. Section through the Moine thrust zone in the region of Kinlochewe, after J. Phemister 31 .

Horizontal scale, 1:60,000, vertical scale, 1:40,000, M = Moine; d = mylonite gneiss, Lewisian; Tn^2 = Applecross series (Torridon); 1, 2, 3, = basal quartzite, "pipe" rocks, and Cambrian fucoid layers.

and overlain by Pliocene, Pleistocene, and contemporary deposits.

In the extreme southwest part of the block, on the islands of Islay and Colonsay, rocks of the Lewisian and Torridon complex appear at the surface, separated from Dalradian rocks by a huge fault.

We shall not dwell here on the description of the early formations developed on the aforementioned islands. It is noteworthy, however, that the Lewisian is represented by orthogneiss; the local rocks of Torridon age have already been briefly described previously.

A. Lower structural stage of the Grampian Highlands Block

The lower structural stage of the Grampian Highlands block is composed of an undisturbed layer of rocks, to all appearances devoid of any signs of inner nonconformity or intrusions, and consisting of an upper part of the Moine series and the full section of the Dalradian series. Only in the extreme southeast of the Grampian Highlands is the sequence supplemented by deposits of the Ordovician (Arenig and Caradocian), apparently, separated from each other and from the Cambrian (upper Dalradian) by faults.

The Moine series developed only in the north and central parts of the Grampian Highlands, where it forms the core of the huge anticlinorium striking northeast. The Great Glen fault partially cut off the northwest flank of the anticlinorium. Considerably better preserved is the southeast flank of this large folded structure of mixed rocks of Dalradian and partly Ordovician age; it is cut short by the Border fault.

The Moine and Dalradian series consist mainly of sedimentary rocks which, in many places, maintain their primary structure. Although the Moine series is relatively older than the Dalradian, they were both simultaneously enveloped by Caledonian regional metamorphism. J. Anderson [10, 11] pointed out that, according to observations in various regions of the Grampian Highlands, both series were deposited in complete conformity. The Moine and the Dalradian rocks are extremely different from each other lithologically. The Dalradian sequence is a very motley gamut of sedimentary rocks, but in the composition of the Moine series, just as is observed in the Northern Highlands of Scotland, the chief rock types are quartzitic-feldspathic granulite and pelitic varieties, which have usually been converted to micaceous schist.

The most widely distributed rocks of the Moine series in the Grampian Highlands are granulite, bearing a great number of local names (see Table 1). This granulite, in the opinion of J. Anderson [10, 11], is stratigraphically equivalent to the upper psammatic group of the Moine series in the Northern Highlands. Higher in the sequence lie micaceous schist and quartzite which in some regions of the Grampians are contemporaneous with the Dalradian complex, and in others with the Moine. According to J. Anderson, the micaceous schist and quartzite on which the basal limestone of the Dalradian lies quite conformably, should be classified as a "transitional group" located in the upper part of the Moine series. The transitional group of the Grampian Highlands is the stratigraphic analogue of the upper pelitic group of the Moine series in the Northern Highlands.

The section of the transitional group of the Moine series is not everywhere the same in various parts of the Grampian Highlands, because of the increase and decrease in the

thickness of the quartzite layers. As J. Anderson [11] believes, the transitional group before metamorphism was composed chiefly of clay with numerous lenses of sandstone at various stratigraphic levels. Even in the small area of the Lochaber region, according to the observations of W. Hardy [22], the thickness of the quartzite layers is variable; quartzite grades laterally into micaceous schist. In places, for example, in the region of Loch Oo in Scotland, just as in northern Ireland [9, 19], pelitic rocks of the transitional group are greatly enriched with a carbonaceous substance.

In the Moine series cross-bedding has been preserved in places, providing data for the study of sedimentary environment in which sediments of the Moine series were deposited. The thickness of that part of the Moine series exposed in the Grampian Highlands is not reported in the literature. Nevertheless, judging from geological maps and sections of different parts of the Highlands (for example, by J. Anderson [11]), the thickness of the granulite of the upper psammitic group (local names are Central Highlands granulite, Aylda flagstones, Struan flagstone, and others) is estimated to be a few kilometers. The corresponding figure for the transitional group (Rannoch and Speyside pelitic schist and quartzite, Lochaber pelite and quartzite, Mull of Oa phyllite, Mull and Fintich quartzite, etc.) is close to 1 km.

Without presenting here in detail a description of the metamorphism which is clearly expressed in the rocks of the Moine series, it is noteworthy that in the northern Grampian Highlands, as W. Kennedy [27] pointed out, a zone of massive, stratified injections of granitic and pegmatitic rocks occur in a layer of sedimentary rocks that have been metamorphosed to migmatite. The whole complex here is completely analogous to that observed in the central Northern Highlands.

The stratigraphy of the overlying Dalradian series also is not uniform for the entire Grampian Highlands. There are a great number of stratigraphic units having a variety of local names. In a number of cases, the sequence of the individual units in the section has not been determined conclusively. This problem was aired in some detail by H. Reed and A. MacGregor [35]. The Dalradian stratigraphy was first worked out for Perthshire by A. Geikie [18]. His scheme, with a few changes, is accepted by the majority of English geologists. J. Anderson [10] recently proposed a new, rational variant of this scheme, in which the individual stratigraphic units of the section are designated by the leading rock type (see Table 1).

The Dalradian series is divided into two parts. The lower Dalradian consists of three groups -- basal carbonate, quartzite, and carbon.

The basal carbonate group, which is traceable from the east coast of Scotland to the west coast of Ireland, generally consists of pure limestone and dark schist. An extremely distinctive detrital layer, the lower part of which consists of disordered rubble as much as 90 cm thick, composed chiefly of Na-granite and cemented by carbonate, occurs in the upper part of the group, and is exposed for 430 km in many regions located between Banffshire and Donegal. Higher in the section the carbonate material decreases due to the gradual increase in silica enrichment ([10], p. 14). By all indications, this detrital layer is not typical glacial deposition (tillite) but sedimentary beds of the marine-glacial type. The rubble probably dropped from floating ice blocks during melting. The layer of rubble is completely conformable both with the underlying limestone and with the overlying quartzite.

The local names of the individual layers of the basal carbonate group in the central part of the Grampian Highlands are: Blair Atholl, limestones, Schiehallion detrital layer; for the regions of Loch Oo and Islay Island -- Islay limestone and Portaskaig detrital layer; for the region of Ballachulish and Lochaber -- Ballachulish limestone and schist.

The second unit from the top of the lower Dalradian, as its name indicates, consists principally of quartzite, but in some places (Islay, Ballachulish) dolomite appears in the group. Local names of the group are: Central Highlands quartzite, Islay quartzite, and Appin dolomites.

The carbon group consists of pelitic rocks containing a considerable amount of scattered graphite (black schist, Isdale clay schist, Appin phyllite, and others).

The thickness of all three groups of the lower Dalradian, according to the cross-section of E. Bayley [35], can be estimated at 2 to 3 km and greater, although this amount is probably not uniform for all sections of the Grampian Highlands.

It has long been known that there is a rhythmic stratification in the dark schist of the carbon group in Perthshire, and cross-bedding in rocks of the quartzite group ([35], p. 27). H. Reed pointed out that evidence of the original sedimentation rhythms were preserved even after the full recrystallization of the rocks; for example, in garnet-staulorite-andalusite schist. Judging from the

Table 1

Correlation of the series and groups of the lower Paleozoic and Precambrian in the Grampian and Northern Highlands of Scotland by J.G.C. Anderson [10]

	Names of stratigraphic subdivisions	Lithology	Grampian Highlands			Northern Highlands
			Central	Loch Oo and Islay Island	Ballachulish and Lochaber	
Ordovician	Caradocian? Arenig Fault	Conglomerate, arkose, and limestone Black siliceous and clayey schist, spilite	Upper series Lower series, (Aberfoyle)	-- --	-- --	-- --
Upper Dalradian Conventional Cambrian	Upper psammitic group	Arkose with seams of pelitic and carbonaceous rocks, "green layers"	Leny arkose, Ben Ledi arkose, "green layers"	Loch Avich clay schist and arkose, "green layers"	--	--
	Upper pelitic and carbonaceous group	Pelite, arkose, and basal lava	Aberfoyle schist, Pitlochry schist, Loch Tay limestone	Tayvallich schist and lava, Tayvallich limestone	--	--
	Lower psammitic group	Arkose, quartzite, and semi-pelitic rocks	Ben Lui arkose and schist	Crinan arkose	--	--
	Lower pelitic and carbonaceous group	Pelite, thin seams of limestone and quartzite	Ben Lawers schist	Ardrishaig phyllite, Shairay limestone	--	--
Lower Dalradian	Carbon group	Graphitic pelite	Ben Igach black schist	Isdale schist	K'yuayl [Kyle?] Bay schist, Appin phyllite	--
	Quartzite group	Quartzite, in places dolomite	Quartzite in Central Highlands	Islay quartzite	Appin dolomite and quartzite	--
	Basal carbonate group	Detritus layer. Limestone and dark schist.	Shiehallion detritus layer. Blair Atholl limestone.	Portaskaig detritus layer, Islay limestone	Ballachulish schist and limestone	--

Table 1 (continued)

	Names of stratigraphic subdivisions	Lithology	Grampian Highlands			Northern Highlands
			Central	Loch Oo and Islay Island	Ballachulish and Lochaber	
Moine (= Torridon)	Pelitic and quartzite transitional group	Pelitic beds with layers of quartzite and, more rarely, limestone	Rannoch and Speyside pelitic schist and quartzite	Mull of Oa phyllite, Mool and Fintich quartzite	Lochaber pelite and quartzite	Upper pelitic group
	Upper psammitic group	Psammites, calcareous rocks in thin seams	Struan flagstone, granulite of the Central Highlands	--	Aylda flagstone	Upper psammitic group
	Striated and pelitic group	Pelite, semi-pelite, and striated rocks Psammite	--	--	--	Striated and pelitic group
	Lower psammitic group	Psammite	--	--	--	Lower psammitic group
	Sub-Moine	Psammite, pelite and hornblende schist	--	--	--	Isolated in the region of Morar. Probably older or younger Lewisian "Windows"; in some of them the rocks are of undetermined stratigraphic position.
	Lewisian	Orthogneiss and paraschist, including graphitic schist and marble	--	--	--	

description of these rocks, they represent a double component flysch in which rhythmic alternation of sandstone and argillite sedimentation occurred before metamorphism.

The age of the lower Dalradian, consisting of the three groups, has been determined in recent times as "probably Precambrian," Precambrian, or late Precambrian [10, 24]. O. Holtedahl [24], and later, J. Anderson [10], pointed out quite convincingly that a clearly defined peculiarity of the late Precambrian sequence, both in Scandinavia and in Scotland, is the presence of glacial and marine-glacial deposits with rubble of Na-granite. In the British Isles, as in Scandinavia, quartzite occurs above the rubble layer. Below the rubble layer, limestone occurs in both areas.

The upper Dalradian, according to J. Anderson [10, 11], part of the Cambrian rests conformably on the underlying rocks of the lower Dalradian carbon group. The Cambrian sequence (upper Dalradian) encompasses (see Table 1) four groups (from the bottom): lower pelitic and carbonaceous, lower psammitic, upper pelitic and carbonaceous, and upper psammitic.

The lower pelitic and carbonaceous group is composed of a sequence of varied pelitic rocks with thin seams of limestone and quartzite. Included are the so-called Ben Lawers schist of the central part of the Grampian Highlands, the Ardrishaig phyllite, and the Shairay limestones (in the region of Loch Oo and Islay Island). The lower psammitic group consists of interbedded arkose, quartzite, and various semipelitic rocks (Ben Lui schist and arkose, and Crinan arkose). Besides the sedimentary rocks (pelite, arkose, and limestone) within the upper pelitic and carbonaceous group, there is a large quantity of basal lavas and tuff (Loch Tay limestone, Pitlochry and Aberfoyle schist in the Central Highlands; Tayvallich limestone, Tayvallich schist and lava). The upper psammitic group is composed of arkose and graywacke with layers of pelitic rocks and thin seams of limestone (green rocks, Ben Ledi arkose, Leny limestone and schist). The fauna discovered by J. Pringle [33] in the Leny limestone on the southern edge of the Grampian Highlands appears in this part of the section. Tribolites of the Pagetia type and other organic forms are suggestive of a Middle Cambrian age for the upper psammitic group of the upper Dalradian.

The Crinan arkose of the lower psammitic group, the Tayvallich and Aberfoyle schist (upper pelitic and carbonaceous group) show indications of rhythmic sedimentation ([35], p. 27). A rhythmic repetition of beds is also characteristic of the arkose, quartzite and graywacke of the upper psammitic group. The thickness of the rhythmic cycle ranges from a few tens of centimeters to 2 m and over, but the usual figure is 0.60 to 1.20 m. Sandstone with fine shingle (1 to 2 cm in diameter) is at the base of the cycle. Higher in the section, the amount of shingle diminishes, the dimensions decrease, and the sandstone gradually has a clay component. The border lines between the rhythmic cycles are sharply defined ([8], p. 485). Cross-bedding in clastic rocks of the upper Dalradian is rarely observed. Against the background of the predominantly gray rocks of the Dalradian series, the Aberfoyle schist (upper pelitic and carbonaceous group) are distinguished in places by their red or purple color. Numerous purple seams are also common in overlying rocks of the upper psammitic group among the greenish-gray and gray arkose and schist ([37], p. 217). Local faulting and nonconformable sedimentation have been discovered ([35], p. 21) in parts of the Dalradian sequence; for example, in the Banffshire district.

It is noteworthy that certain components of the upper Dalradian sequence, for example, the Loch Tay layer of limestone, at the base of the upper pelitic and carbonaceous group, were deposited under astonishingly uniform environmental conditions and are traceable along the strike for hundreds of kilometers, through all the Grampian Highlands and as far as northeast Ireland.

The effusive rocks of the upper pelitic and carbonaceous group are extremely significant. Spilitic pillow lavas are predominant, sometimes associated with tuff and volcanic agglomerate [35]. Felsite, keratophyre, and granite porphyry are also present among the effusive rocks. "Green layers," which lie at the base of the upper psammitic group, are highly distinctive both in outward appearance and in composition (they have an abundance of epidote and chlorite). They are metamorphosed volcanic ash or sedimentary rocks that have developed through the disintegration of magmatic rocks [10].

Thanks to the findings of J. Pringle, the middle Cambrian age of the upper psammitic group of the upper Dalradian has been conclusively demonstrated. This circumstance has permitted J. Anderson to tentatively date the entire upper Dalradian series as Cambrian [10].

The thickness of the Cambrian (= upper Dalradian) is considerable, and by adding thickness figures reported for individual sections by various authors) it is possible to estimate the thickness of the Cambrian as 3 to 4 km [8, 12, 35, 37], and in places

even more.

Ordovician beds are uppermost in the section of sedimentary and volcanic deposits of the lower structural layer of the Grampian Highlands. Ordovician deposits occur in a narrow area along the southeast border of the Grampian Highlands, in the zone of the Border fault.

The Arenig is represented chiefly by spilite, which in places has preserved the structure of pillow lavas. Subordinate to the spilite are layers of dark clay, flinty schist, and seams of volcanic ash. The thickness of the Arenig formations has been accurately determined on the island of Arran where it reaches 300 m. Its relationship to the Cambrian is not entirely clear; supposedly the Arenig is separated from the Cambrian by a fault. A typical silicic-volcanic formation of the Arenig is distinguished by a clearly defined unconformity separating it from the overlying deposits which are tentatively dated as the Caradocian layer of the Ordovician. The Caradocian is composed of coarse breccia and conglomerate, and is overlain by arkose and clastic limestone. The thickness of these deposits is not great -- about 60 m [8, 35]. The intrusions associated with the lower structural layer of the Grampian Highlands fall into two groups -- "ancient" and "young." Greenstone rocks and the granite belong to the category of "ancient" intrusions [35].

The greenstone rocks combine blanket deposits and laccoliths. They are usually gabbro, more rarely enstatite gabbro, with which is associated ultrabasic and anorthositic rocks. All these rocks have been metamorphosed and have been changed, respectively, to epidiorite, amphibolite, or hornblende schist and serpentinite.

It is significant that the largest stratified bodies and an important laccolith (Ben Vrackie) of the basic rocks occupy a distinct place in the profile of the lower structural layer of the Grampian Highlands, and occur in the upper part of the lower Dalradian series and in the lower half of the Cambrian sequence (upper Dalradian), or, in other words, between the quartzitic and the upper pelitic and carbonaceous groups. Spilite, keratophyre and porphyry, with which are also associated numerous intrusions of basic rocks of the upper pelitic and carbonaceous group of the upper Dalradian, occur stratigraphically above these intruded bodies in the upper Dalradian.

"Ancient granite" is the name given [35] to a group of silicic rocks that were intruded either at the time, directly afterwards, or before the regional metamorphism of the Moine and Dalradian rocks. Certain intrusions were consolidated before the surrounding rocks (augen-granite) were sheared. The main part of the ancient granite is involved in the regional folding. Large bodies of similar granite are not common; the intrusions have taken the general form of small conformable bodies, veins, and seams penetrating the surrounding rocks. According to the observations of J. Barrow [35], the areas of highest metamorphism of sedimentary rocks (sillimanite zone) are also associated with the zones of intrusion of the ancient granites. Other zones of metamorphism occur in order of decreasing intensity outward from the periphery of separate centers of magmatic activity. The principal types of ancient granite, according to J. Barrow, are: biotitic gneiss, bimicaceous granite and gneiss, and oligoclase-biotite-muscovite and oligoclase-muscovite gneiss. Ancient granite intrusions in the opinion of H. Reed, are especially abundant in that section of the upper and lower Dalradian sequence that is located between the quartzite of the Central Highlands and the layer of limestone at Loch Tay. Here, in the opinion of the writer, it is necessary to remember that the greatest part of the blanket deposits of the basic type are also associated with this section of the Dalradian series.

D. Reynolds [35] considers the "ancient granite" of J. Barrow to be migmatite, contemporaneous with the Caledonian orogeny, and as emerging in connection with the advancement of the fronts of migration of Na-Ca-Si and K-Fe-Mg-Al. The porphyroblasts of albite and oligoclase developed in schist and gneiss, in his opinion, during the course of migration of the elements through pelitic rocks initially deprived of the albite molecule. Of all these migrating elements only sodium came from the depths of the interior.

Later came the extremely abundant and varied "Caledonian intrusions" or "young granite." Here, late Silurian and relatively younger lower Devonian intrusions [35] are conditionally distinguished. Both groups of intrusions are younger than the regional metamorphism affecting rocks of the Moine and Dalradian series.

Gabbro and granite are common rock types of the late Silurian intrusions. Gabbro forms huge stratified bodies in the northeast Grampian Highlands. The intrusions include: olivine gabbro, troctolite (forellenstein), noritic gabbro, olivine norite, hypersthenic gabbro, and peridotite. In places, there are quartzitic diorite, syenite, and granite. The well-known contaminated rocks -- norite with cordierite and garnet -- are in the zones of contact of the intrusions with

surrounding rocks of the argillite type. Around the intrusions, aureoles resulting from thermal, contact metamorphism have developed distinctly, in the limits of which andalusite and cordierite are developing in the schist and chert formation is occurring.

Late Silurian granite forms a number of small bodies as well as huge masses. The form of some of the masses is curious, for example, the Cairngorm is nearly a tabular blanket. The most extensive rock type of this group is biotitic granite containing small amounts of muscovite and microcline. The smaller bodies are composed of hornblende granite, tonalite, diorite, and aplite. Dike rocks of the same age group (aplite, pegmatite, quartz-porphyry, microgranite, porphyrite, various types of lamprophyres and peridotite) are abundant and varied. In the zone of contact of the granite with the surrounding rocks, the phenomena of thermal, contact metamorphism (resulting in the development of chert and skarn) are clearly expressed.

The lower Devonian intrusions are clearly developed especially in the southeast Grampian Highlands. The largest intrusion, Glen Etive, is formed by a complicated circular complex of four or five granite intrusions, which were intruded one after the other in connection with four or five crater-shaped settlings [7, 35]. A notable circular intrusion, Ben Nevis, is also connected with a trough of subsidence and is mixed with highly differentiated granite and quartzitic diorite [6, 12, 35]. Silicic rocks, granitic diorite, the Moor of Rannoch, form the intrusion of Cruachan. The intrusions in the vicinity of Loch Lomond, which consist of ultrabasic, basic, middle, and silicic rocks, are completely differentiated. The numerous dikes and sills, especially concentrated in the southwest part of the Grampian Highlands, where entire systems of dikes belong to the same age group. The intrusion of most dikes along fissures occurred after the development of the first circular intrusions of granite in the zones of caldera subsidence. The dikes are composed of felsite, lamprophyre, and various porphyrites.

Regional metamorphism clearly affected the rocks of the Moine and Dalradian series over the whole extent of the Grampian Highlands. As early as 1893, J. Barrow [35] pointed out that in the southeast part of the Highlands the following zones of metamorphism are distinguishable from the Border fault northward: 1) clastic mica (very narrow zone), 2) beginning of neogenic mica, 3) biotite, 4) garnet, 5) staurolite, 6) cyanite, 7) sillimanite. These zones, in the opinion of J. Barrow, form a gigantic aureole around the intrusions of "ancient granite." C. Tilley [35] put together a similar map of the various zones of metamorphism for the southwest part of the Highlands, but for index minerals he selected chlorite, biotite, and almandite. The metamorphism of the southwest Highlands was interpreted again in 1923 by E. Bailey, who suggested the following zonal division: 1) fine, weakly developed mica, 2) mica in appreciable quantity, 3-a) mica with garnet, 3-b) mica with albite. In comparatively recent times (1946), W. Kennedy published a new map of the metamorphic zones of the Grampian Highlands [27] on which he showed a sillimanite zone comprising the northern part of the Highlands, and a comparatively narrow cyanite zone on the periphery of the first zone; further, to the south and southwest, are zones of garnet, biotite, and chlorite.

The metamorphic zonation, regardless of the means of representation and selection of index minerals, without doubt exists objectively, but it has been, unfortunately, insufficiently related to the stratigraphy of the Moine and Dalradian series and to their tectonics. "Metamorphism in the various regions of the Highlands," writes A. Harker ([5], p. 189-190), "over the larger part may be studied independently of the complicated and difficult questions of stratigraphy and tectonics." This singular approach, involving the isolated study of metamorphic phenomena which has existed up to the present, has not aided in solving a number of important questions on the geology of Scotland. Among such questions, for example, is that of the interrelationships during the time of regional metamorphism and folding. E. Bailey [12] considered that metamorphism and folding occurred practically at the same time. J. Ellis and C. Tilley [16] have insisted on a wide development of inversions of the metamorphic zones. H. Reed [35] supports the opinion that regional metamorphism took place later than the chief folding movements.

The tectonics of the lower structural layer of the Grampian Highlands is still far from being interpreted in terms of the generally accepted ideas of the stratigraphy of the Moine and Dalradian series. At the present time, there are few attempts at interpretation of the tectonics of the Grampian Highlands. J. Barrow [35] has advanced an idea about a structure of a "concertina" (harmonic) type. J. Green [19-21] considered that the tectonics of the southwest part of the Highlands is determined by simple synclines and anticlines. There are other more widespread views on the tectonics of the Grampian Highlands, all based on the ideas of E. Bailey [12]. In his opinion, in the south and southwest parts of the Highlands a series of large horizontal folds has developed, complicated by fractures of a special kind

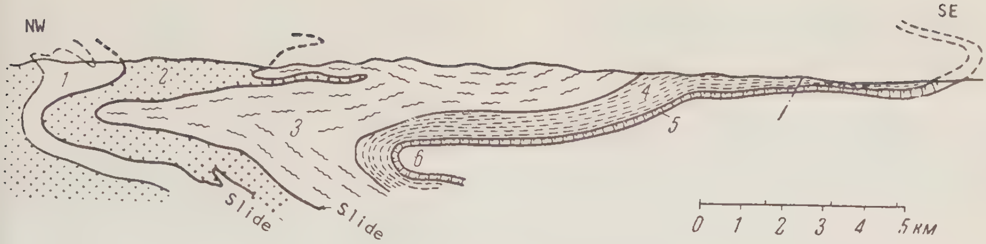

FIGURE 5. Pattern of the folded structure of the southwest part of the Grampian Highlands, after E. Bailey (in section). After H. Reed and A. MacGregor 35 .

1-6 Various stratigraphic elements of the Dalradian section.

(slide, and fold faulting) and by components of certain tectonic blanket deposits. Secondly, these primary structures were crumpled into a system of open or compressed synclines and anticlines, and in places into secondary recumbent folds under the influence of superposed folding stresses. The predominating strike of the folds is northeast (see Fig. 5).

The views of E. Bailey are supported, developed, and modified by many investigators of the geology of Scotland -- H. Reed, E.M. Anderson, J. Ellis, C. Tilley.

B. The upper structural stage of the Grampian Highlands block

The upper structural stage of the Grampian Highlands block is many-layered. As was already noted previously, in the structure of the upper stage, Devonian, Carboniferous, Permian, Triassic, Pliocene, Pleistocene formations, and contemporary deposits are all involved. The present description will be limited to a brief characterization of only the Devonian formations.

Lower Devonian lava and sedimentary rocks occur in a small area located on the northwest rim of the Grampian Highlands in the volcanic plateau of Lorne which lies northwest of Loch Oo, and in the region of Glencoe and Ben Nevis; the tectonic relationship of these rocks is extremely unusual. On the plateau of Lorne [35], Lower Devonian rocks include andesite and felsite lava, agglomerates, tuff, ash, and also sedimentary rocks -- conglomerate, sandstone, arkose, shale, and limestone in thin seams. A sedimentary-igneous rock layer of lower Devonian age lies nonconformably on the metamorphic schist of the Dalradian series. Thin sedimentary rocks (30 to 60 m thick) lie at the base of the section. The overlying lava, which is interbedded with pyroclastic rocks, forms a layer more than 600 m thick. The thickness of all layers of sedimentary and igneous rocks decreases to the east in the Grampian Highlands. Fossils of fish, euryterids, millepedes, ostracods, and also of plants occur in the sedimentary rocks.

In the region of Glen Coe [35] the thickness of the sedimentary-igneous series reaches 1200 m. Here, the sedimentary rocks are also subordinate. The lower Devonian formations lie nonconformably on metamorphic schist and make up a zone of caldera subsidence bounded by a circular fault. Besides andesite, rhyolite is also present in the lava complex. In analogous conditions, Lower Devonian formations form the core of a crater-shaped trough of subsidence at Ben Nevis [6, 35]. In the opinion of English geologists, the eruptions of Lower Devonian lavas in all the cases described were associated with the action of volcanoes of the central type and occurred in subaerial conditions.

Rocks of Middle Devonian age occur only in the northern Grampian Highlands, in the region of the southern coast of the large bay of Moray. The local section of Middle Devonian rocks is analogous to a similar section on the northwest coast of the same bay, briefly described on a previous page. The Middle Devonian consists of a thick layer of sedimentary rocks (conglomerate, sandstone, shale, and subordinate thin seams of limestone containing many fossils of fish and, in places, of plants). There are few igneous rocks (andesite) in the Middle Devonian sequence. The sedimentary rocks lie nonconformably on various older rocks of the Moine and Dalradian series.

The upper Devonian occurs in a considerably smaller area in the same region of the southwest coast of Moray Firth. Sedimentary breccias occur in the lower part of the section and are commonly overlain by sandstone with seams of clay, limestone, and marl, in which there are numerous fish

fossils. In the sandstone, well-defined cross-bedding is common. The Upper Devonian sequence lies nonconformably on both rocks of Middle Devonian age and on metamorphic schist of the Moine and Dalradian series.

The Devonian rocks are relatively flat-lying, but their occurrence is partly restricted by faults.

* * * * *

In concluding the description of the separate large blocks forming the Scottish Highlands, it should be called to mind that the Grampian Highlands are bounded on the southeast by the Border fault. Farther southeast lies a great depression -- Midland Valley -- an area where sedimentary rocks of Devonian and Carboniferous age are extensive. On the downthrown southeast side of the Border fault, rocks of the Upper Silurian (Downtonian, [8]) are involved; these are unknown in the Grampian Highlands. The Downtonian is composed of sandstone, tuff, and argillite with fossils of fish and arthropods. The apparent thickness of the deposits reaches 730 m. Thick conglomerate, sandstone, and igneous rocks of the Lower Devonian rest conformably on rocks of Downtonian age.

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ON THE PROCESSES OF NEPHELINIZATION AND AEGIRIZATION OF PYROXENITE AND THE ORIGIN OF ALKALINE ROCKS OF THE IOLITE-MELTEIGITE TYPE

by

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Structural interrelationships of nepheline and pyroxene in alkaline rocks of the iolite-melteigite type, and also the peculiarities of the geological environment of these rocks attest to the postmagmatic origin of nepheline. The author comes to the conclusion that the formation of iolite-melteigite, which forms complex masses of ultrabasic-alkaline rocks together with pyroxenite, is explained not by the intrusion of alkaline magma but by metasomatic change -- by nephelinization and aegirization of other ultrabasic rocks.

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Geological studies of recent years connected with prospecting for ores of niobium, zircon, rare earth minerals, and titanium, have led to the exploitation of new types of economic mineral deposits, the most important of which are carboniferous deposits. Their most characteristic peculiarity consists in a close spatial and genetic association with rather rare alkaline rocks -- nonfeldspathic nepheline-pyroxene rocks of the iolite type [3]. Iolite and its melanocratic and leucocratic variants (melteigite and iolite-urtite) are most commonly found together with typical ultrabasic rocks (pyroxenite and peridotite, more rarely olivinite) and in combination they take massive oval or round shapes, occurring, as a rule, in the more stable parts of the Earth's crust (ancient platforms and shields). The structure of such masses commonly manifests a specific regularity wherein alkaline rocks of nepheline-pyroxene composition form a peripheral circular zone around a central cell of ultrabasic rocks. This generally simple zonality can be complicated as a result of the emergence of other rocks, for example, melilitic, micaceous or apatitic rocks, which also form circular or incomplete circular zones. In the central part of masses of ultrabasic (alkaline) rocks, carbonates, sections of which are circular in form, also occur; they commonly have a concentric zonal structure.

The presence of concentric zonality in these masses is explained by the majority of investigators as a result of gradual magmatic intrusion, the composition of which has changed because of magmatic differentiation. Moreover, the following general sequence of intrusional activity is generally assumed: 1) ultrabasic phase -- pyroxenite, peridotite, and other rock types; 2) alkaline phase -- melteigite, iolite, and other rock types; 3) carbonate phase and other rocks that combine with them (apatitic-magmatite, siliceous carbonate, and others).

The criteria generally used in determining the sequence of magmatic intrusion are the following: 1) interlacing of some rocks by veins of other rock types; 2) presence of some "primary" lineations and "flow textures"; 3) change of composition and texture of "earlier" rocks on contact with "later" rocks. Although these criteria usually permit determination of the relative age of some rocks, they are completely inadequate for a well-defined conclusion about their magmatic origin. This is especially true of the genesis of alkaline rocks of nepheline-pyroxene composition. These rocks, in combination with pyroxenites are the leading components of the plutons in which we are

nterested, establishing their name as "plu-ons of ultrabasic-alkaline rocks."

Iolite and its variants (melteigite, iolite-ırtite, etc.) may be viewed as a bimineral eries of rocks, the composition of which lepends on the quantitative relationship of nly two minerals -- nepheline and pyroxene. n this series, according to Brogger, may be listinguished: melteigite with a nepheline con-ent of less than 50 percent, iolite with a epheline content of 50 to 70 percent, and ırtite, which contains over 70 percent epheline [7]; jacupirangite is commonly in-luded as the final member of the series nder review. In this connection, the follow-ng points are noteworthy. A common feature f all members of the series, according to rogger, is the combination: nepheline and reen pyroxene low in aluminum oxide (gen-rally aegirite-diopside, aegirite being the ninor component). If this definition is trictly adhered to, then neither jacupiran-ite -- an augitic rock -- nor urtite -- an egirite-nepheline rock -- can be put in a ontinuous series with nepheline-pyroxene ocks of the iolite-melteigite class. This ircumstance has already been noted by amsay in a description of urtite (aegirite-olite) from the Lovozero pluton [9].

Similar to jacupirangite, essentially an ugitic, (generally non-nepheline rock) urtite ombines with geologically distinct bodies nd preserves all the basic peculiarities of omposition and texture (for example, urtite n the Lovozero alkaline mass). At the same ime, investigators who have studied iolite nd melteigite from ultrabasic-alkaline rocks ote that these rocks are extremely altered n composition and texture. If any section of ne zone of alkaline rocks generally corre-ponds in its quantitative mineral content to ny member of the class jacupirangite-ırtite, then, within separate exposures, and ven within lumps of ore, an irregular equence of essentially leucocratic and nelanocratic varieties of rocks are present. Moreover, even in urtitic rocks richest in epheline, aegirite commonly is not present.

No less remarkable is the circumstance nat within each pluton, pyroxene from the one of alkaline rocks is analogous in its asic chemical properties to the pyroxene rom the zone of pyroxenes; i.e., in a plu-on where the central zone is composed of iopside pyroxenite, the pyroxene from the one of alkaline rocks also contains diopside rith some aegirite; it is exactly so in plu-ons where pyroxenite of augitic composition ave developed, and in alkaline rocks augite ontaining a small amount of aegirite is resent, etc. In other words, although the omposition of pyroxene in alkaline rocks efinitely depends on the composition of the pyroxene in the ultrabasic rocks, there is no relation between the amount of aegirite in the pyroxene and the total alkalinity of the rock. The degree of alkalinity is determined chiefly by the amount of nepheline in the rock. Thus, the difference between rocks within the iolite-melteigite series amounts to a change in the relative content of nepheline.

If we also take into account that very commonly, the zone of alkaline rocks has no clear "intruded" contacts with an adjoining pyroxenite zone, then it is apparent that the question of the magmatic origin of alkaline rocks is similar to the question of their rela-tion to pyroxenite. Moreover, the following problems are of special significance: 1) the peculiarities of composition and texture of the rocks in the zone of contact of pyroxenite and nepheline-pyroxene rocks; 2) textural interrelationship of nepheline and pyroxene in iolite-melteigite; 3) paragenesis of alkaline pyroxene and pyroxene of ultrabasic rocks. No work in the fairly extensive literature on massive ultrabasic-alkaline rocks exists which treats these problems in any detail. Therefore, pyroxenite and nepheline-pyroxene rocks, which in the opinion of the majority of investigators are independently intruded, are regarded as practically independent of each other.

As has been noted earlier, in all works devoted to the study of ultrabasic-alkaline rocks, it is assumed that ultrabasic rocks form earlier than alkaline rocks. Inasmuch as alkaline rocks form generally peripheral sections of the plutons, then, assuming the magmatic origin of these rocks, the emer-gence of circular faults around a core of ultrabasic rocks and the intrusion through them of alkaline magma must be postulated. Where extensive erosion has obliterated the contact between ultrabasic and alkaline rocks, such an assumption, at first, seems probable. Nevertheless, our observations in studying zones of alkaline rocks in a number of plu-tons, along with the accounts of certain other investigators [1, 4, 6, 10], show that the geologic position of the indicated rocks in the plutons observed, and their structural-textural peculiarities exclude the possibility of intrusion of alkaline magma of "nepheline-pyroxene composition" which crystallizes, forming the corresponding rock. The sup-position seems more probable that nepheline-pyroxene rocks emerge as the result of the postmagmatic change of ultrabasic rocks, first of pyroxenite, caused by the develop-ment of nephelinization and aegiritization [2, 3].

Admitting, as a general rule, that nepheline-pyroxene zones are present on the outer, peripheral part of the plutons, it is necessary to mention that these rocks can

also be found among the central sections of ultrabasic rocks. For an example, one can point to the Ozernaya Varaka pluton, studied by V.A. Afanas'yev [1].

As follows from the schematic geologic map of this pluton (Fig. 1), nepheline-pyroxene rocks which are chiefly in the peripheral part, are also found in the midst of pyroxenite. Thus, essentially pyroxene variants of nepheline-bearing rocks -- "melteigite" -- occur directly in the zone of pyroxenite. As the data of V.A. Afanas'yev show, the chief mineral of pyroxenite and nepheline-pyroxene rocks is the same -- titanium-bearing augite. Megascopically, melteigite essentially differs from pyroxenite only by the presence of separate crystals of nepheline in a basic mass including pyroxene. The question arises, does melteigite actually develop independently of pyroxenite as a result of the crystallization of a new magma of alkaline composition? Additional studies of the pluton in 1955 show that an extremely unequal distribution of nepheline in the mass of pyroxene crystals is present in the rocks of Ozernaya Varaka.

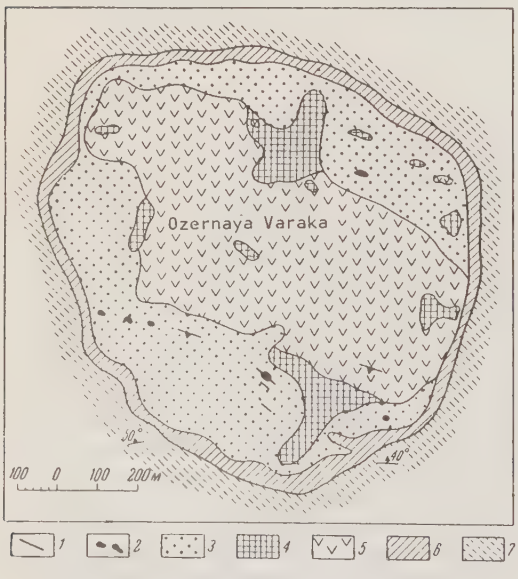

FIGURE 1. Geologic map of the pluton of alkaline rocks of Ozernaya Varaka, after V.S. Afanas'yev

1-Cancrinitic syenite; 2-urtite; 3-iolite; 4-melteigite; 5-pyroxenite; 6-contact rocks; 7-granite gneiss.

In turning from the separate peripheral sections, composed of fine-grained nepheline-pyroxene rocks, generally of one type, to the central zone of pyroxenite it is possible to see that no sharp contact exists between these rocks. Within the transitional zone in the inner mass of pyroxenite vein-like, lense-like, or irregularly-distributed impregnations of nepheline are present. In places in the pyroxenite monomineral nepheline veins and lenses occur. Sections of the rock which contain a small amount of nepheline phenocrysts ("melteigite") do not differ megascopically from pyroxenite. With an increase in the amount of nepheline phenocrysts, the outward appearance of the rock changes and "melteigite" replaces "iolite." Maximum saturation of pyroxenite by nepheline produces a rock type that may be formally defined as urtite. As V.A. Afanas'yev remarks, they "are found only in the form of small islets in iolite" [1].

Analogous relationships between pyroxenite and nepheline-pyroxene rocks are characteristic of a number of other plutons. Thus, for example, in the description of these rocks in the Afrikand pluton, P.N. Chirvinskiy, M.S. Afanas'yev, and S.G. Ushakova [6] state that "The width of the nepheline pyroxenite belt is subject to strong fluctuations. . . No clear contact between the nepheline pyroxenite and the more centrally located fine-grained pyroxenite was observed. Generally, there was a barely noticeable transition manifested by the gradual disappearance of nepheline. Isolated spots of nepheline pyroxenite were found outside of this zone, in fine-grained pyroxenite. In the southern part of the pluton the nepheline pyroxenite is penetrated by numerous pegmatitic veins of schorlomite-nepheline composition. The main rock mass has a massive structure, but nepheline and apatite, which are part of the composition of the rock, have collected in isolated seams, in places, giving a striated appearance to the rock."

From the description above it follows that the transition from ultrabasic rocks to the zone of silicic rocks, and variations in rocks of nepheline-pyroxene composition are both controlled by the behavior of one mineral -- nepheline, which develops both monomineral separations and zones of "impregnation" in pyroxenite. It is apparent that the chief component of these and other rocks -- pyroxene -- has developed from a single process of crystallization. Therefore, the explanation of the development of nepheline-pyroxene rocks would be aided by the determination of the time of development of nepheline in relation to pyroxene. The following solutions to the paragenetic relationships are possible: either the nepheline crystallized more or less simultaneously with the pyroxene from the same magma, or nepheline is an epigenetic mineral in relation to pyroxene.

In the first case nepheline-pyroxene rocks must be viewed as a phase variation of pyroxenite. Inasmuch as the crystallization of intruded magma must have developed from the periphery of the intrusion to the center, then, the nepheline-pyroxene rocks, which occupy the peripheral part of the intrusion and are in direct contact with the surrounding rocks, must have crystallized somewhat earlier than the pyroxenite. Moreover, there is the supposition that the alkalis -- the most mobile components of a rock melt -- were not distributed regularly throughout the whole volume of the crystallizing magma, but, on the contrary, have become localized only in the peripheral part of the intrusion.

In the second case, nepheline must be a postmagmatic mineral, developing as a result of the displacement of pyroxene during the action of later solutions on the emplaced rock. If this supposition is correct, then nepheline-pyroxene rocks should be regarded as metasomatically changed -- nephelinized -- pyroxenite. Nepheline would then be a later mineral in relation to pyroxene, and nepheline-pyroxene rock would differ in its textural peculiarities from a typical intruded rock.

The study of sections of nepheline-pyroxene rocks from various plutons of Karelia, Kola Peninsula and Siberia confirms the correctness of the second supposition. As will be shown below, in textural relations the rocks under discussion are closer to metamorphic than to magmatic types and are characterized by the appearance of singular corrosion textures and structures arising as the result of the process of nephelinization of pyroxenite [2, 3].

The morphological, textural peculiarities of this type are more clearly apparent in the initial and middle stages of the process of nephelinization, before a regular displacement of pyroxene by nephelite has occurred over the whole mass of rock. Therefore, it is quite clear how monomineral branching "veins" or irregular islands of nepheline grains, corroding the pyroxene crystals, are present among the pyroxene aggregates which wholly preserve the texture of the primary rock (Fig. 2). If the nepheline occurs as relatively small xenomorphic precipitations, commensurable in size with the relics of the pyroxene crystals, then, in proportion to the development of replacement, they form aggregates of a nepheline-pyroxene composition with an allotriomorphic, granular texture, which is reminiscent of the granoblastic texture of metamorphic rocks (Fig. 3). Numerous relic sections of pyroxenite and separate larger crystals of pyroxene occur among these aggregates (Fig. 4). In the more nephelinized varieties, pyroxene islands enclosed in a mass of even-grained rocks occur on the site of the relic sections.

Where the process of nephelinization has taken place with extreme intensity, a more or less even-grained nepheline-pyroxene rock has emerged. It should be emphasized that, in such places, an "abrupt contact" may be observed between the primary pyroxenite and the nepheline-pyroxene rock (Fig. 5).

The corrosion textures have a different character where nepheline has crystallized in the form of idiomorphic, isometric grains. Commonly, fine relic inclusions of pyroxene are abundant. Therefore the texture of the rocks may be called poikiloblastic or screen-like (Fig. 6). The idiomorphic grains of nepheline commonly contain only a small number of relic inclusions of pyroxene. In this respect the corrosion texture may be erroneously taken as the subhedral, granular texture of a magmatic alkaline rock (Fig. 7). Nevertheless, even among the idiomorphic grains of "early" nepheline there may be observed uniformly orientated relics of larger pyroxene crystals; this quite obviously is an affirmation of the late development of nepheline in comparison with pyroxene (Fig. 8).

Study of the relationships between nepheline and pyroxene show that nepheline-pyroxene rocks of the iolite-melteigite type develop through the metasomatic change of pyroxenite. From this conclusion it is possible to assume that in individual relic structures of primary pyroxenite, zones of alkaline rocks will occur. The correctness of this supposition is supported in a whole series of plutons of ultrabasic-alkaline rocks. In some of them, for example, in the Odikhincha (Oegincha) pluton, in the central section of the "iolite" zone, numerous centers of normal pyroxenite, isolated from each other and covering an area of some tens of meters, are present. The possibility of finding relic structures of pyroxenite should always be considered in mapping iolitic and melteigitic "intrusions" where independent zones of pyroxenites are lacking. Thus, for example, in the Gulinskiy pluton, where a number of investigators assume the intruded nature of melteigite, I discovered monomineral sections in the midst of alkaline rocks and also sections of apatitized pyroxenite, commonly found in plutons where independent zones of these rocks have begun to be mapped. E.L. Butakova indicates, though with some reservations, the possibility of the metasomatic genesis of nepheline-pyroxene rocks of the Gulinskiy pluton and of the metasomatic character of the nepheline in them, in her latest work. She writes: "Nepheline

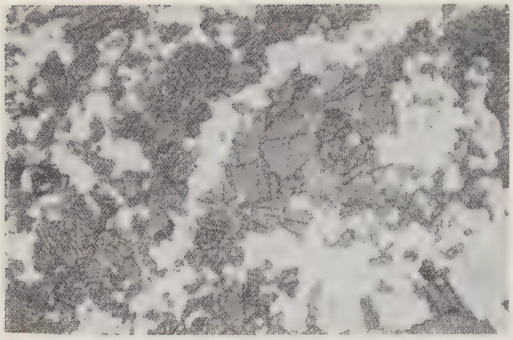

FIGURE 2. Initial stage in the process of nephelinization.

Nepheline (light) develops veins or separations amid the aggregates of pyroxene grains (dark). Without analyzer, enlargement 25X.

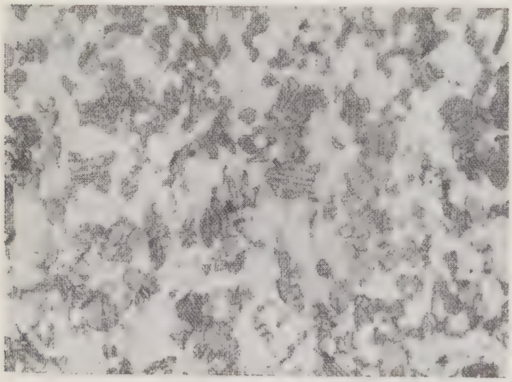

FIGURE 3. Granoblastic texture of nephelinized pyroxenite ("iolite").

Nepheline - light, pyroxene - dark. Without analyzer, enlargement 25X.

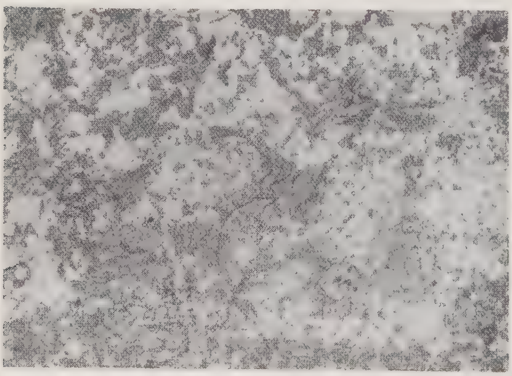

FIGURE 4. Corrosional texture of nephelinized pyroxenite ("melteigite").

In the basic mass of Nepheline-pyroxene composition, the remnants of the primary pyroxenite are included.

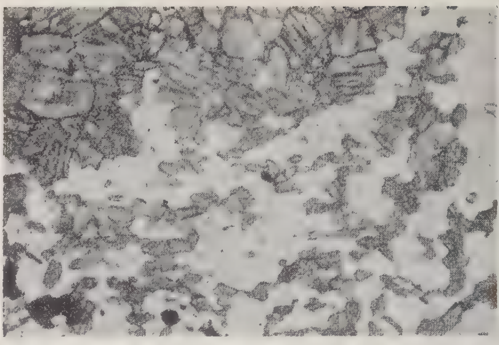

FIGURE 5. Contact of pyroxenite and a nepheline-pyroxene rock of the iolite type.

The pyroxene in both rocks is of the same type (not aegiritized). Without analyzer, enlargement 25X.

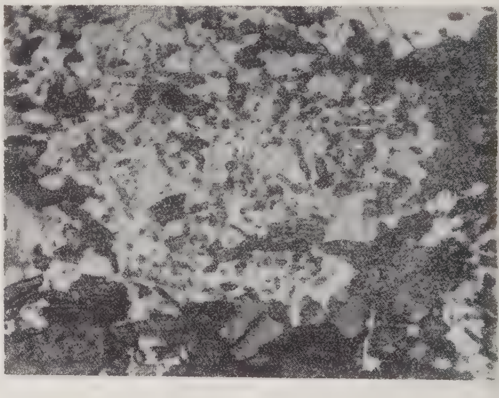

FIGURE 6. Corrosional poikiloblastic texture of a nepheline-pyroxene rock.

Individual, larger grains of the nepheline (central part of photomicrograph) are full of small "microlites" -- remnants of pyroxene crystals. Without analyzer, enlargement 50X.

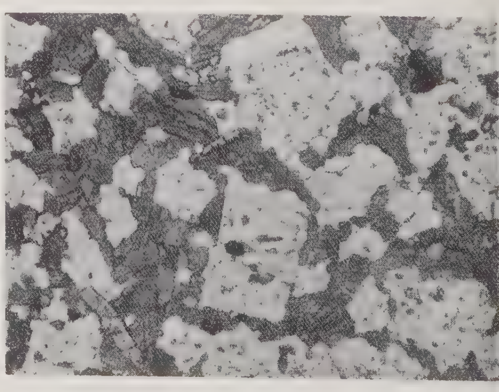

FIGURE 7. Corrosional poikiloblastic texture with idiomorphic grains of nepheline (light).

Without analyzer, enlargement 25X.

behaves not like a mineral that has crystallized directly from a magma, but it apparently develops even in intruded rocks in considerable measure, if not entirely, by metasomatic means" [4].

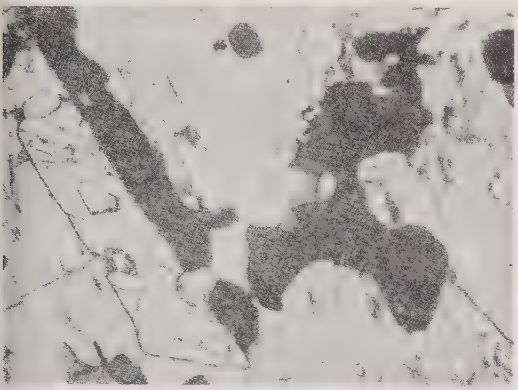

FIGURE 8. Corroded counterpart of pyroxene-phenocryst in a larger grain of nepheline. One element of a twin structure (black), appearing under polarized light like two independent grains, is under extinction. Enlargement 50X.

In the preceding discussion one important circumstance was not considered, i. e., aside from dependence on the quantitative and textural relationship of nepheline and pyroxene in nepheline-pyroxene rocks, these rocks differ from typical pyroxenite in that the pyroxene in them, as a rule, contains a certain amount of segirite and has a greenish color. Inasmuch as such alkaline pyroxene (aegirite-diopside or aegirite-augite) is a basic component of iolite or melteigite, this circumstance is considered one of the arguments for the genesis of these rocks from especially alkaline magma. Moreover, the presence of predominantly pyroxene sections down to monomineral sections taken from zones of alkaline rocks is explained by the compositional changes of crystallizing magma, to which various types of rocks in the urtite-jacupirangite class correspond.

In permitting such an explanation, a fact already mentioned is ignored -- the discrepancy between the total change of alkalinity of the rock and the preservation of a more or less uniform (though small) aegirite content in the pyroxene. Let us recall that even in nepheline-pyroxene rocks, where the amount of nepheline is 70 to 80 percent or more, aegirite usually does not exceed 15 percent of the total pyroxene composition. It also seems to be forgotten that nepheline-pyroxene rocks develop in some plutons where, with a nepheline content of 50 percent or more, the pyroxene approaches the composition of common diopside or augite, differing in no way from the pyroxene of primary pyroxenite of the pluton. We have observed individual sections of such rocks, for example, in the Odikhincha pluton in Siberia.

In this connection it seems interesting to note also the presence of small amounts of alkaline pyroxene in the midst of the usual pyroxenite which form the central parts of the zonal masses. In structural relationship, these rocks, which occur in the pyroxenite zone beyond its line of contact with the alkaline rocks, differ in no way from normal pyroxenite surrounding them and therefore may easily be passed over. Only in microsections of such varieties of pyroxenite can it be established that the pyroxene contains aegirite and is analogous to pyroxene from a nepheline-pyroxene zone.

In the light of these facts one characteristic feature of the pyroxene of iolite-melteigite, its zonality, acquires a different significance. As many investigators remark, for pyroxene from iolite-melteigite the characteristic zonal coloring comes from the fact that the central part of the crystals is composed of colorless or light-colored diopside or augite, and the peripheral part is green pyroxene containing a small amount of aegirite [5, 8]. The peculiarities of such zonality are the lack of any clear boundaries between zones and the spotty coloring (Fig. 9). In places, it can be seen that the zonality is directly related to the process of change in the pyroxene, in particular its change to mica. In individual cases, it may be seen that the "greening" of pyroxene, caused by the appearance of aegirite, occurs not only in the peripheral part of the crystal but also in cracks in its central part (Fig. 10). All this brings forth the assumption that the alkaline nature of pyroxene in zones of nepheline-pyroxene rocks is caused by secondary change, the process of aegiritization. Aegiritization is part of the metasomatic transformation of primary pyroxenite and accompanies nephelinization. However, as already pointed out [3], aegiritization can also take place independently of nephelinization, preceding it. This explains the fact that aegiritized pyroxenite ("jacupirangite") occurs within the pyroxenite zone, beyond the contact with nepheline-pyroxene rocks.

Aegiritization shows up with special clarity in the contacts of pyroxenite and veins bearing rocks of alkaline composition (nepheline-garnet, ziolite and others), which intersect the pyroxenite beyond the zone of nepheline-pyroxene rocks. In this case "greening" of the pyroxene crystals, caused by aegiritization, is observed in bands roughly parallel to the contact of the pyroxenite and the alkaline vein. This band "intersects" the variously orientated pyroxene crystals adjoining the alkaline vein

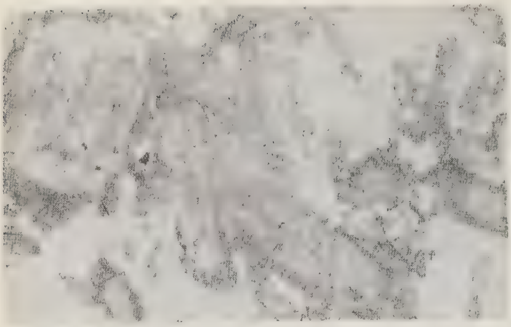

FIGURE 9. Zonal coloring ("greening") of pyroxene in nephelinized pyroxenite.

Aegiritized sections of pyroxene in the photograph have a dark coloring. Without analyzer, enlargement 50X.

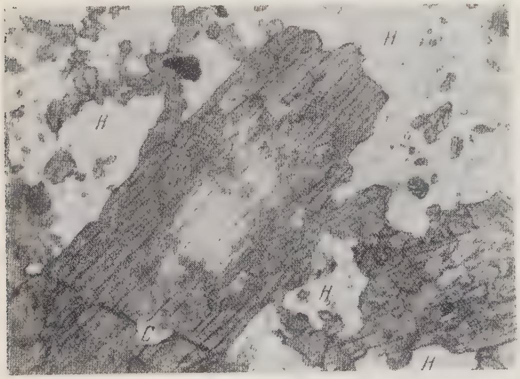

FIGURE 10. Zonal coloring of pyroxene crystals.

Aegiritized sections, which have a green color, are dark in the photograph. Relics of primary pyroxene (colorless) are concentrated in the central part of the grains, H-nepheline, C-mica (phlogopite).

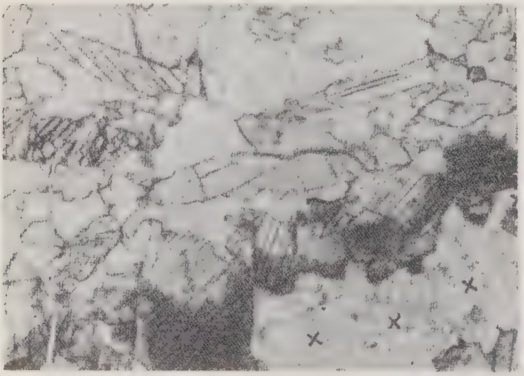

FIGURE 11. Band of aegiritization -- "greening" (dark in the photograph), occurring along the contact of the pyroxenite (upper half of photograph) and a vein of alkaline rock (denoted by crosses). Without analyzer, enlargement 50X.

(Fig. 11). The sections of crystals outside the band preserve the usual coloring of pyroxenite.

Summarizing all of the above, we come to the following fundamental conclusion.

The formation of alkaline rocks, which determines the zonal structure of the masses we are discussing, was caused not by processes of magmatic differentiation and the intrusion of particular batches of "alkaline magma," but by the metasomatic change of pyroxenite and other originally magmatic ultrabasic rocks. In this connection, nepheline-pyroxene rocks of the iolite-melteigite[1] type should not be incorporated in a single group with aegirite-nepheline rocks of the urtite type (aegirite-iolite) nor should all these rocks be related to one continuous series resulting from differentiation of alkaline magma.

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[1]Jacupirangite, in our opinion, is also not a special intruded rock crystallizing from alkaline magma, but a weakly nephelinized pyroxenite.

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THE NEPHELINIZATION OF PYROXENITE AND MARBLE

by

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The nephelinization of pyroxenite and marble is regarded by the author as an original contact process. Under the action of alkaline intrusions of urtite-iolite into the pyroxenite, nephelinized pyroxenite, melteigite and iolite emerge. The evolution of nepheline to marble with the formation of silico-carbonaceous rocks has been observed at the exocontact of an alkaline intrusion with the xenoliths included in it and especially clearly so on the border.

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The history of the problem

Great attention has been paid to the process of nephelinization in the last decade, particularly in the explanation of the origin of the alkaline complex of rocks in the Haliburton-Bancroft region (Ontario), of certain varieties of alkaline rocks in the mass of Alno (Sweden), urtite and iolite-urtite in the Khibinskiy mass (Kola Peninsula), and in a number of other places.

The term "nephelinization" was used by W. Gummer and S. Burr [4] in explaining the origin of nepheline-bearing rocks on account of the essentially carbonaceous deposits of the Grenville series of the Haliburton-Bancroft region. At the same time H. Eckermann [3] regarded nephelinization as the concluding stage of the process of phenitization occurring during the action of carbonaceous magma on acid rocks of the surrounding Archean strata in the Alno region. Both of these authors and, later L. Moyd [5], and I.P. Tikhonenkov described the development of a neogenic nepheline chiefly from feldspar, noting corrosion of the dark colored minerals only during the most intense stage of this process.

S. Strauss and F. Truter [6] used the term "nephelinization" in a different sense -- as the process of development of nepheline through pyroxenite. E.L. Butakova [1] refers to the conversion of pyroxene to nepheline and to the development of nepheline at the point of contact with marble at the Gulinskiy intrusion.

Analogous phenomena are discussed in the present article, i.e., the development of nepheline in pyroxenite and marble, observed in contact zones of urtite-iolite intrusions located in the basin of the Balyktyg-Khem River in southeast Tuva.

The nephelinization of pyroxenite

The Dakhunur urtite-iotite intrusion is as yet the only one in the Tuva alkaline province where nephelinization of pyroxenite has been observed.

The intrusion lies nonconformably amid marbles of upper Proterozoic age and is confined to a zone of intense crushing in the central part of an anticlinal structure. Rocks of the urtite-iolite class form two steeply dipping bodies having an oval shape resulting from contemporary erosion and shearing. The largest of them -- East-Dakhunur -- does not exceed 2 km in length with a width of 1 km (Fig. 1, I). The urtite-iolite rocks are spatially connected with pyroxenite in earlier stages of development and have an active effect on them in the zone of contact, as the result of which rocks which approach melteigite in composition evolve.

Brief characterization of contact rocks. Iolite-urtite and urtite from the contact zone have the appearance of compact, coarse-grained, meso- or leucocratic rocks of a gray color, quite uniform in composition and texture. They are composed chiefly of

nepheline (for optical constants see Table 1), which averages about 85 percent in urtite, 65 percent in urtite-iolite, and of titanium ferroaugite, the amount of which increases from one percent in urtite to 13 percent in iolite-urtite.

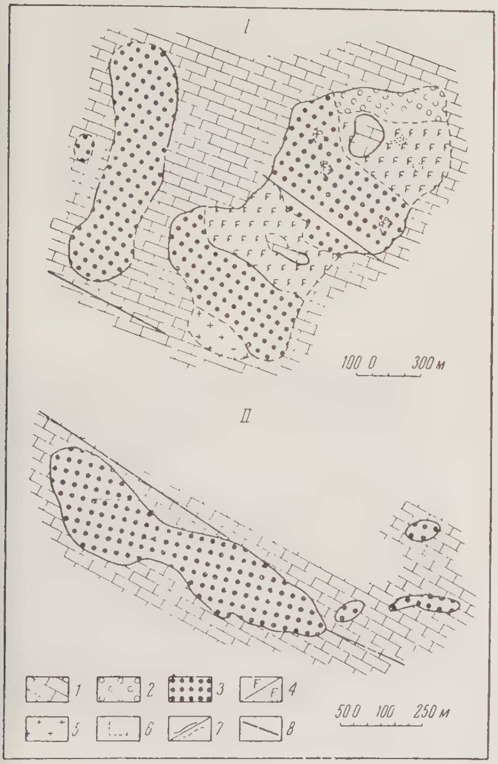

FIGURE 1. Plan of the structure of the Dakhunur (I) and Chiksk (II) urtite-iolite intrusions.

1-marble of Proterozoic age; 2-nepheline syenite; 3-rocks of the urtite-iolite class; 4-pyroxenite (1), nephelinized pyroxenite (2); 5-plagiogranite; 6-area of maximum extent of converted xenolith of marble; 7-granite. (1) original, (2) coarse fault-block screes; 8-faults.

Nepheline and pyroxene decrease with the development of secondary minerals, among which we note cancrinite, plagioclase, calcite, garnet, and hornblende. The accessory minerals are represented by apatite, more rarely by sphene and titanium magnetite. The texture of the rocks is pan and hypidiomorphic granular.

Unaltered pyroxenite is represented by fine- or medium-grained, compact, black or greenish-black varieties of quite stable composition and uniform texture. These are almost monomineral rocks chiefly of titanium ferroaugite, the optical constants of which are given in Table 1. Due to the postmagmatic replacement of pyroxene by garnet (up to 7 percent), feldspar (up to 15 percent), epidote (about 2 percent), calcite (about 5 percent), and sometimes even plagioclase (about 2 percent) the amount of pyroxene decreases appreciably -- to 60 or 65 percent. The accessory minerals are commonly sphene, titanium magnetite and apatite. The latter rarely increases above 2 or 3 percent. The texture of unaltered pyroxenite is pan hypidiomorphic granular.

The geologic conditions and pattern of development of nephelinization. Nephelinization occurred chiefly at the exocontact of an alkaline intrusion. Because of extensive erosion it is difficult to judge the magnitude of the given process. Observations in separate sections along a series of exposures of rocks in situ show that the zone of nephelinization apparently was quite considerable (up to 50 m). The rocks changed in varying degree, beginning with the appearance of individual nepheline grains in the pyroxenite, through metasomatic melteigite to iolite with poikiloblastic sections of nepheline.

In two cases, nephelinization of pyroxenite was noted at a considerable distance from the visible contact. In particular, at the peak of Dakhunur mountain (near the triangulation mark) there is a small section of nephelinized pyroxenite separated from the contact by approximately 200 m. It is quite possible that the surface of the contact was uneven and at the given spot an alkaline intrusion lies under the pyroxenite.

Included in the intruded iolite-urtite, near its contact with pyroxenite, xenolithoid sections of pyroxenite are found from 15 cm to tens of meters in diameter; they are commonly formed by the breaking up of large blocks. In places small, highly-resorbed spots and elongated insertions of pyroxenite with indistinct and broken edges (Fig. 2) are included in the intruded iolite-urtite. As a rule, in the above case and in more distinctly zenolithoid sections, retarded nepheline penetrates into the pyroxenite.

The first stages of nephelinization of pyroxenite are registered by the appearance of fine grains and very thin (1 to 2 mm) veins very difficult to discover in hand specimens. Farther away these fine precipitations are grouped and expand in the form of shapeless spots clearly visible on the surfaces of the lumps. Nepheline commonly accumulates in the cracks.

In outcrops and individual specimens it is possible to see the various stages of the process of nephelinization. In Figs. 3 and 4

Table 1

Optical constants of the chief rock-forming minerals

Optical constants	Titanium ferroaugite from: Pyroxenite	Titanium ferroaugite from: Iolite-urtite	Titanium ferroaugite from: Nephelinized pyroxenite	Nepheline from: Iolite-urtite	Nepheline from: Nephelinized pyroxenite
2V+	62 - 64°	63°	Not determined	--	--
N_γ: [001]	51 - 58°	53°	"	0°	0°
N_γ' (N_ω)	1,741 - 1,733	1,735 - 1,731	1,735 - 1,733	1,538	1,538
N_α' (N_ϵ)	1,720 - 1,716	1,721 - 1,717	1,718 - 1,746	1,533	1,535
Pleochroism N_γ	Brown with violet shades			None	
Pleochroism N_β	Light brown with violet shades				
Pleochroism N_α	Greenish brown				
Texture	Hyp- and pan-idiomorphic granular		Poikiloblastic	Hypidiomorphic granular	Poikiloblastic

are shown two of them, namely: local development of nephelinized pyroxenite (Fig. 3) and formation of uneven-grained melteigite with relic sections of partially converted pyroxenite (Fig. 4). In the well-advanced process rather large-grained melteigite and iolite appear. Pyroxenite nephelinized in varying degree occur over an area without any apparent regularity. The process of nephelinization was accompanied by corrosion and partial displacement of pyroxene by nepheline. The nepheline appeared at first in small amounts, developing skeletal forms of growth. At the same time, similar precipitations of nepheline, seemingly isolated (Fig. 5), have a single optical orientation, and characterize the initial growth stage of an individual, large poikiloblast. In the final stage of nephelinization, as a rule, coarse poikiloblastic sections developed, up to 0.5 cm in diameter (Fig. 6), in which there were more than ten seemingly-fused grains of pyroxene undergoing a partial conversion to hornblende and garnet. Especially interesting is the inter-penetration of nepheline and pyroxene, accompanied by the disintegration of the latter mineral into a number of simultaneously fading sections (Fig. 7), the edges of which have fantastically sinuous outlines due to corrosion. The presence of corrosional and poikiloblastic textures is the outstanding mark of metasomatic iolite and melteigite.

Features of chemical affinity. Nephelinized pyroxenites are fairly simple in their composition, being a mixture of nepheline and pyroxene, which, according to the optical constants, are the same as the minerals in alkaline rocks and pyroxenite (Table 1). Nevertheless their composition is extremely variable and fluctuates, even within one specimen of standard dimensions, particularly for nepheline, which ranges from single precipitations up to 30 percent. In individual cases, garnet of the andradite-grossularite class appears in nephelinized pyroxenite, sometimes up to 10 percent.

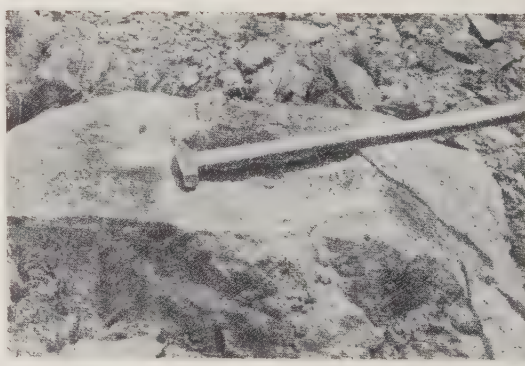

FIGURE 2. Resorbed sections of pyroxenite in urtite.

In explaining the basic features of the process of nephelinization of pyroxenite, special attention was paid to chemical composition and both unaltered pyroxenite (see Table 2, anal. 1) and pyroxenites converted in varying degree were analyzed, beginning with pyroxenite with a minor nepheline content (up to 10 percent) (see Table 2, anal. 2)

lown to melteigite with a nepheline content of 30 to 35 percent (see Table 2, anal. 3).

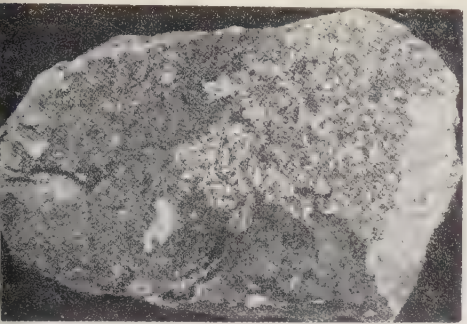

FIGURE 3. Nephelinized pyroxenite. Wedge-shaped formation.

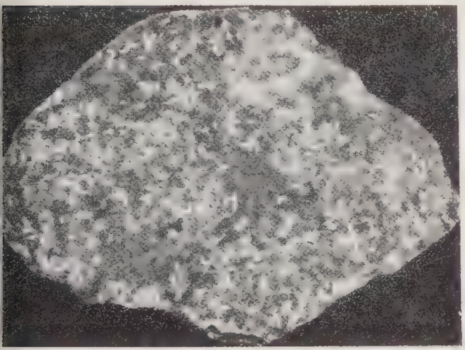

FIGURE 4. Uneven-grained melteigite with remnants of lightly converted pyroxenite.

Comparison of the chemical analyses of inaltered and slightly nephelinized pyroxenite hows the active role of the alkali even at he beginning stage of nephelinization. At he same time, attention is called to the oss of a comparatively large amount of 'eO. In other respects, except for some ncrease in Al_2O_3 and decrease of CaO, the omparative analyses are close.

If we compare the analysis of nephelinized yroxenite corresponding to melteigite in omposition (see Table 2, anal. 3), to that f a more intensive stage of nephelinization, t should be emphasized that this same tenency in the change of chemical content is reserved, but it is more sharply displayed. 'he content of alkali and aluminum increases narkedly, iron is lost in considerably larger mounts (about one half) than CaO (less than /3) and MgO. At the same time there are ppreciable amounts of CO_2 in nephelinized ocks.

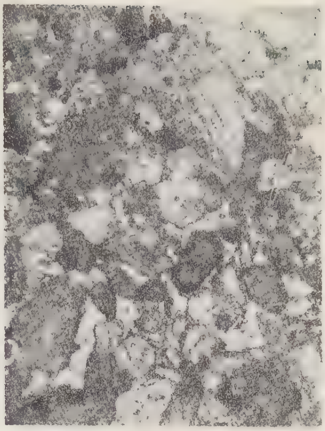

FIGURE 5. Skeletal forms of growth of poikiloblastic nepheline. Enlargement 20X, with analyzer (nepheline = H).

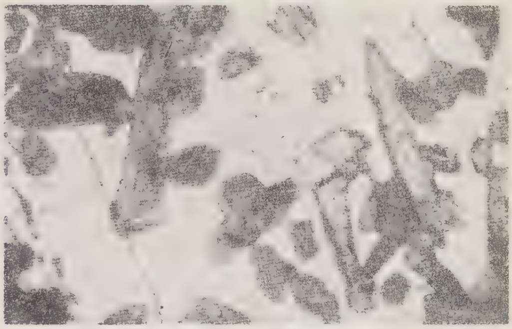

FIGURE 6. Poikiloblastic precipitations of nepheline (light). Enlargement 20X, with analyzer.

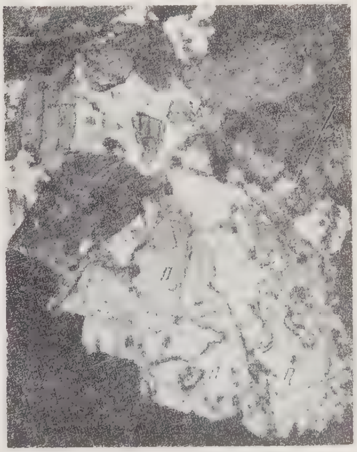

FIGURE 7. Penetration of nepheline into pyroxene along the joints. Enlargement 46X, with analyzer. Isolated sections of a single grain of pyroxene are denoted by the letter P.

Table 2

Chemical composition of nephelinized pyroxenite

Oxides	Pyroxenite		Slightly nephelinized pyroxenite		Nephelinized pyroxenite of melteigite composition	
	Sample 1		Sample 2		Sample 3	
	Weight %	No. of electropositive ions in a standard nucleus	Weight %	No. of electropositive ions in a standard nucleus	Weight %	No. of electropositive ions in a standard nucleus
SiO_2	41,36	416	41,30	423	39,56	387
TiO_2	1,24	9	1,07	8	1,29	9
Al_2O_3	11,56	137	12,46	150	21,45	248
Fe_2O_3	4,34	33	4,58	35	1,67	12
FeO	11,50	99	8,56	75	6,06	50
MnO	0,25		0,20		0,08	
MgO	5,61	84	5,96	90	4,17	61
CaO	22,82	246	21,29	234	16,35	172
Na_2O	0,82	16	2,64	53	5,78	110
K_2O	0,04	—	0,49	6	1,34	16
H_2O^-	0,12	53	0,02	24	0,06	62
H_2O^+	0,68		0,32		0,89	
S	0,03		0,19		0,04	
F	0,02		0,00		0,00	
Cl	0,00		0,00		0,03	
CO_2	0,00		1,12	15	0,80	10
BaO	0,00		0,00		—	
Total	100,39	1093	100,20	1113	99,57	1 137
Analyst	V. Nekrasova				E.I. Lomeyko	

Note: Comma represents decimal point.

Using the oxygen method for computing the analyses and in determining the migration of matter in the nephelinization process, we get the following values for judging the addition and loss of the elements:

Addition			Loss		
111	ions	Al	74	ions	Ca
94	"	Na	49	"	Fe^{+2}+Mn
16	"	K	29	"	Si
11	"	OH	23	"	Mg
3	"	C	21	"	F^{+3}

The recorded peculiarities of chemical composition are not accidental, since a magma solution has acted on the pyroxenite -- having fused, more quickly than the whole urtite compound, with a fairly high percentage of alkali and Al_2O_3. The SiO_2 urtite and pyroxenite are quite close in amount. Rocks of the urtite-iolite group are often called the basic alkaline rocks. With close correlations of silica in the contacting rocks, this group has not played a decisive role in the nephelinization process, because its deficit or surplus has not been large.

Apparently, in the pyroxenite-urtite contact zone an exchange takes place with the active elements. To the pyroxenite is added chiefly Al_2O_3 and Na_2O, and the released iron, CaO, and MgO are lost to the adjacent sections of urtite. As a result, melteigite emerges in the contact zone which is formed, on the one hand, from pyroxenite in consequence of the addition of nepheline to them, and on the other, from urtite in consequence of the increase of pyroxene (titanium ferroaugite) in them.

Nephelinization of Marble

The urtite-iolite intrusions of Tuva [2] known at present occur in marble of upper Proterozoic age, but only in one of them -- Chikskaya -- can a clear nephelinization of marble be observed.

The Chikskaya intrusion fits into a large regional fault which forms the limb of an anticline. The rocks of the urtite-iolite group form four steeply dipping bodies which lie conformably to the intruded marble and have numerous xenoliths included in them

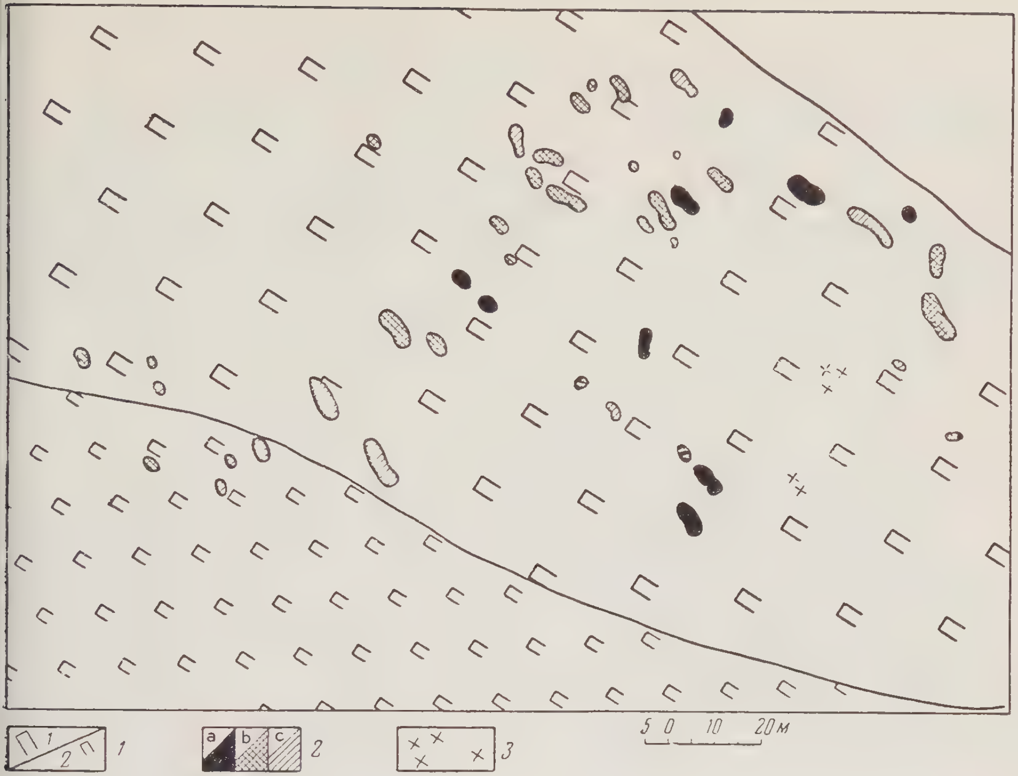

FIGURE 8. Plan of distribution of altered xenoliths containing marbles. Chikskaya intrusion.

1. Streaky, banded complex of iolite-urtite and iolite (1-pegmatoid and coarse-grained variety; 2-medium-grained variety).
2. Marble xenoliths, metamorphosed in varying degree (content of silicate minerals: a=70-50%; b=50-20%; c=20-5%).
3. Sites of xenolithic marble with dimensions less than 2m.

FIGURE 9. Marble xenolith.

Contoured, grayish zone of carbonaceous silicate rocks; white is the unaltered cell of the xenolith

(Fig. 1, II).

Development of nepheline in marble was observed in the contact of the intrusion and especially clearly at the border with the xenoliths.

The xenoliths of marble have various dimensions and, as a rule, are oriented in the direction conforming to the foliation of the intruded rocks. An idea of the form and dimensions of their bed is given in a sketch (Fig. 8) which we made for the small, most eroded section in the western part of the Chikskaya intrusion (Fig. 1, II).

Marble xenoliths, of which the largest is shown in Fig. 9, are always metamorphic.

In the peripheral zone, of varying thickness (on the average about 1 to 2 m), they have been transformed into unusual carbonaceous silicate rocks. The content of silicate minerals, in which pyroxene (of the augite-hedenbergite class), apatite and nepheline are found, varies from 5 to 70 percent, not

Table 3

Chemical composition of nepheline and pyroxene (in %)

Oxides	Nepheline		Pyroxene	
	Iolite-urtite	Carbonaceous silicate rock from xenolithic marble	Iolite	Carbonaceous silicate rock from xenolithic marble
	anal. 1	anal. 2	anal. 3	anal. 4
SiO_2	41,44	41,40	43,71	44,46
TiO_2	none	none	1,67	1,28
Al_2O_3	34,34	34,31	6,12	5,46
Fe_2O_3	0,57	0,12	4,45	5,54
FeO	none	0,19	15,93	17,63
MnO	none	none	0,36	0,39
MgO	0,11	0,05	4,30	2,72
CaO	1,12	1,18	22,04	21,05
Na_2O	14,46	15,96	1,21	1,08
K_2O	6,42	5,34	0,10	traces
H_2O^-	0,18	0,13	not found	0,18
H_2O^+	0,42	0,73	0,29	0,30
CO_2	not det.	0,60	—	0,34
S	—	traces	—	traces
loss on ign.	0,46	—	—	—
Total	99,52	100,01	100,18	100,43
Materials	R. M. Yashina	V. A. Kononova	R. M. Yashina	V. A. Kononova
Analyst	V. A. Kharitonova	K. P. Sokova	E. I. Lomeyko	K. P. Sokova
N_γ : [001]			47°	
2V+			66°	
N_γ' (N_ω)	1,541	1,539	1,739	1,759
N_α' (N_ϵ)	1,534	1,533	1,722	1,729
Pleochromism N_γ	not present		Cinnamon-brown with violet tinge	Greenish-yellow
Pleochromism N_β			Dark brown	Grass green
Pleochromism N_α			Greenish or yellowish brown	Dark grass-green
Plan of absorption			Ng> Nm> Np	Np> Nm> Ng

Note: Comma represents decimal point.

Table 4

CaO and MgO content in enclosing marbles and in an xenolithic carbonate
(Analyst V. A. Akhvonen)

Oxides	Enclosing marbles (average of 3 tests)	Carbonate from	
		Central part of xenolithic marble	Carbonaceous-silicate rock enclosing xenolith
	anal. 1	anal. 2	anal. 3
CaO	51.90	54.12	54.25
MgO	0.56	not observed	0.10

only in various xenoliths but also within one xenolith, and decreases from the periphery to the center.

The silicate minerals penetrating the xenoliths develop first in the interstices between grains of calcium, partly corroding them and gradually displacing them. At the same time, the proportion of alkali in the nepheline composition (Tables 2 and 3) changes in comparison with the same mineral from the iolite-urtite surrounding the xenolith (Table 3, anal. 1), i.e., the sodium content increases and potassium decreases. Pyroxene from unaltered xenolithic marble (Table 3, anal. 4) differs from the titanium ferroaugite of the surrounding alkaline rocks (Table 3, anal. 3) by having an increased content of silica and iron and a decrease in aluminum, magnesium, and calcium.[1]

In the altered xenoliths, besides the new formation of silicate minerals, there occurs an impoverishment and almost complete disappearance of magnesium and an accretion of calcium in the carbonaceous part of the xenolith (Table 4, anal. 1). The element admixtures, as the spectral analyses made by T. S. Reshetina indicate, do not display differences in the composition of carbonates from xenoliths and the intruded layers.

Iolite-urtite near granite with marble also undergo conspicuous changes and within a zone 10 to 15 cm thick are converted to melteigite, in which an increased amount (up to 23 percent) of calcium is present.

The development of nepheline, together with pyroxene and apatite, in xenolithic marble and the related conversion of the surrounding iolite-urtite to melteigite testify to the definite interaction among sharply differing rocks (marble and alkaline rock), which accompanied the displacement of CaO and CO_2 in the intruded rock, and SiO_2, Al_2O_3, alkali, iron, and phosphorous in the xenolith.

Conclusion

The cases examined in this article on the development of nepheline in pyroxenite (Dakhunur intrusion) and marble (Chikskaya intrusion) are situated exclusively in zones of contact of these rocks with urtite-iolite intrusions and may be viewed as a unique contact metamorphic process.

Nephelinization of pyroxenite attracts special interest when accompanied by the development of characteristic poikiloblastic and corrosional textures, brought about by the late appearance of the nepheline, metosomatically displacing pyroxene. Nephelinized pyroxenite corresponding to differing degrees of nephelinization are heterogeneous in composition and present a taxitic texture. If, at the initial stage of nephelinization, slightly altered pyroxenite emerges with some nepheline, then true alkaline rocks corresponding in makeup to melteigite and partly to iolite are developed in the final stage. The only outstanding feature of these rocks in relation to pyroxenite is the presence of nepheline.

It is natural to see the source of the alkaline which caused the development of nepheline in rocks of ultrabasic composition in the alkaline intrusion itself.

[1]The optical properties change slightly simultaneously -- the indices of light refraction are raised, and the greenish tone of pleochromism appear with a reversed plan of absorption (Table 3).

The nephelinization of pyroxenite we studied in the zone of contact with the Dakhunur intrusion has much in common with the phenomena of nephelinization described by B. Strauss and F. Truter [6] in the Spitskop mass (South Africa). From the data of these authors, the pyroxenite on both sides of their veins of iolite, and, in rarer cases, even of urtite, is permeated with nepheline. As a result, wherever the pyroxenite is, rocks emerge which correspond in composition to jacupirangite and iolite.

Let us emphasize an important genetic feature: the action of iolite magma on the pyroxenite led to the formation at Dakhunur, on the other hand, there was wide development of melteigite, i.e., rocks containing more nepheline than jacupirangite, because here the intrusion which acted on the pyroxenite was close to urtite in composition.

The nephelinization of pyroxenite in contact with the urtite-iolite intrusions of Tuva arouses interest because of the common problem of the genesis of individual groups of alkaline rocks, in particular melteigite. In the case we described melteigites show no indications of being primary magmatic rocks, but make up a group of unusual contact-metasomatic formations. The same thing occurs at Spitskop in regard to jacupirangite.

In a number of alkaline intrusions, melteigite and jacupirangite (Kola Peninsula, north Siberian platform) are most often considered as representatives of a magmatic series of urtite-iolite-melteigite-jacupirangite. It must be remarked that melteigite and jacupirangite usually accompany alkaline intrusions of mixed composition, where ultrabasic rocks (pyroxenite and others) are involved. Such an association is apparently not only spatial but even more closely genetic when, as at Dakhunur and Spitskop, the possibility of the appearance of metasomatic alkaline rocks, particularly due to the nephelinization of pyroxenite, is not excluded.

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THE STRUCTURE OF THE MINE FIELD OF THE SLYUDYANKA PHLOGOPITE DEPOSIT

by

K. N. Korzhinskaya

ABSTRACT

New geologic data support the opinion that there are two rock types in the Archean Slyudyanka series -- carbonaceous and gneissic, gathered into isoclinal folds. They also explain the principal factors which contribute to the formation of mica.

* * * * *

Stratigraphy and structure of the phlogopite deposit

The geologic structure of the Slyudyanka phlogopite deposit and the relationship of the phlogopite veins to structural elements have often been discussed in geologic literature.

One of the first investigators of Slyudyanka -- S. S. Smirnov [4] wrote about the tectonics of this deposit: "Regarding the tectonics of the region, it is extremely complex. . . Disregarding the details, one can speak of the existence of certain folds of a northwest strike with axes plunging in the same direction." The crystallized rocks which compose the Slyudyanka region he divided into two series: "One, consisting of primarily crystallized limestone, and the other consisting chiefly of garnet and biotitic gneiss."

Later investigators -- A. S. Suloyev, P. P. Pilipenko, P. V. Kalinin and others [1] -- also noted in the region of the Slyudyanka deposit a complex fold with a principal anticline trending northwest, and dissecting the crystallized strip into a lower gneissic section, composing the arch of the anticline, and an upper section of limestone.

N. N. Padurov and E. P. Shchukina also support the two-part division of the Slyudyanka crystallized series, but, in contrast to the preceding investigators, they consider the limestone as the lower, and the gneiss the upper section.

E. P. Chuykina, F. V. Kuznetsova, and others put a different light on the structure and stratigraphy of the Archean Slyudyanka. They divide the Archean metamorphic Baikal complex into four series: the Sharyzhalgaysk -- Lower Archean; Slyudyansk -- Middle Archean; and Kharagol'sk and Bezymyannyy -- Upper Archean. They divide the Slyudyansk series into three sub-series consisting of 21 stratigraphic units; the lower and upper are predominantly marble, the middle is gneissic. They assign rocks to a stratigraphic unit according to the indication of the frequency of joint occurrence without regard to their mineralogy. E. P. Chuykina and F. V. Kuznetsova regard the Slydyanka deposit as monoclinal, developing as the southwest limb of a large Archean anticlinorium.

This same point of view is held by B. M. Ronenson [3]. Based on the work of N. V. Frolova, "The conditions of sedimentation in the Archean era," he borrows the idea of a cycle of sedimentation formulated by N. B. Vassoyevich [3] in application to flysch and flysch-like strata of coal-bearing deposits and used by M. A. Zavalishin and N. A. L'vova (1954) in relation to Precambrian rocks. He complicated the stratigraphy of the ancient Slyudyanka series even more in the same way. Maintaining the three-part division of the Slyudyansk series, he separated, instead of 21, 16 units, deposited in the course of 16 cycles of sedimentation and composed of 48 layers which are separated into 33 blocks.

Each unit, according to B. M. Ronenson, corresponds to a regular cyclic interchange of four phases of sedimentation in the Archean marine basin: 1) carbo-aluminous, characterized by blocks of pyroxene-

amphibolic gneiss; 2) silico-aluminous, represented by biotitic gneiss; 3) silico-carbonaceous, corresponding to striated quartzitic-diopsidic and quartzitic-dolomitic rocks, and 4) carbonaceous, corresponding to marble. All the blocks form into the belt-like monoclinal structure of the mining zone preserving the uniform thickness and composition of each block.

Thus, there are two contradictory opinions on the stratigraphy and texture of the Slyudyanka deposit: first -- on the presence in the Archean Slyudyanka strata of two series and isoclinal folding, supported by S. S. Smirnov, A.S. Suloyev, P.P. Pilipenko, P.V. Kalinin; and second -- on three sub-series which form a monocline, put forward by E.P. Chuykina, F.V. Kuznetsova and supported by B.M. Ronenson.

As a result of the geological survey of the mining zone area of the Slyudyanka deposit in 1951-1953 at a scale of 1:1000, much new factual material was collected which confirms the earlier view of the complex isoclinal folded structure. The data of the survey also confirmed the presence of only two series in the crystallized belt of the Slyudyanka strata: marble and gneiss.

Inasmuch as the geologists of Slyudyanka had adhered up to the present time to the more recently established opinion on the monoclinal structure of the deposit with a tripartite division of the Slyudyanka series, it seems expedient to us to review once more, on the basis of the new material, the structure and stratigraphy of the Slyudyanka deposit, and also to point out certain regularities in the localization of the phlogopite veins.

The studies of 1951-53 show that the Slyudyanka Archean strata are divided into two series: 1) marble, with a seam of leucocratic, small-flaked biotitic gneiss in the upper part, and 2) gneissic rock of a pyroxene-amphibole, biotitic and biotitic-garnet composition. The structure of the deposit is very complex, characterized by the presence of numerous compressed and open folds, complicated by smaller folds; closely compressed isoclinal folds with strike and dip principally in a northwesterly direction predominate.

A geologic map of the mining zone of the deposit (Fig. 1) reflects the relationship of the marble and gneiss series mentioned above. The gneisses form a strip trending northwest having a width of 300 to 1000 m, in which the principal mines are located. To the east and west of the mining zone marble with seams of dolomite and quartzitic-diopsidic rocks have formed. The marble and the quartzitic-diopsidic rocks in separate lenses and blocks are also noted even within the gneissic belt of the mining zone. In the layer of crystallized marble near the contact with the gneissic belt a row of seams of leucocratic biotitic gneisses occurs -- a type most often found to the west and more rarely to the east of the gneissic belt. This leucocratic biotitic gneiss is even more widely developed within the gneissic belt of the mining zone.

On the basis of geologic cross-sections, constructed from mapping data, and documents of prospecting and test-well core drilling, it has been ascertained that the marble and leucocratic biotitic gneiss which bound the gneissic belt of the mining zone from the east and west, and also the leucocratic biotitic gneiss and marble found within the gneissic belt, comprise a single unit.

The elements of stratification in the periclinal parts of the structure serve as an indication that the marble forms with the block of leucocratic biotitic gneiss the lower part of the crystallized series of the Slyudyanka deposit, and the pyroxene-amphibole gneiss, biotitic, and biotitic-garnet composition form its upper part. The marbles in the whole folded structure of the region form a nucleus of anticlinal folds, and the gneisses fill out the synclinal troughs.

At the base of the stratigraphic cross-section of the Slyudyanka stratum (Fig. 2) lie crystallized limestones of the lower series, the thickness of which has not been established, but N.N. Padurov and E.P. Shchukina estimate it to be 650 m. In the upper part of the layer of crystallized limestones lie lenses of quartzitic-diopsidic rocks with a thickness of 15 to 20 m. Above lies a 15 to 20 meter block of leucocratic small-flaked biotitic gneiss, enriched in places with hypersthene, garnet cordierite, graphite, and sillimanite. Leucocratic biotitic gneiss is covered with a 15 to 20 meter layer of crystallized limestone. The marble in contact with gneissic rocks -- pyroxene-amphibole and small-flaked biotitic gneiss -- has been turned to diopside.

The upper series is composed from the base upward of pyroxene-amphibole gneiss with an average thickness of 30 to 40 m. Above lies hornblende-pyroxene-biotitic gneiss 25 to 30 m thick, which are covered by striated garnet-biotitic gneiss with a thickness of 15 to 20 m, which are replaced above by biotitic-pyroxene gneiss, the thickness of which has not been established.

The shape of the folds is extremely varied: side by side with the wide, and for the most part, open folds of the lower

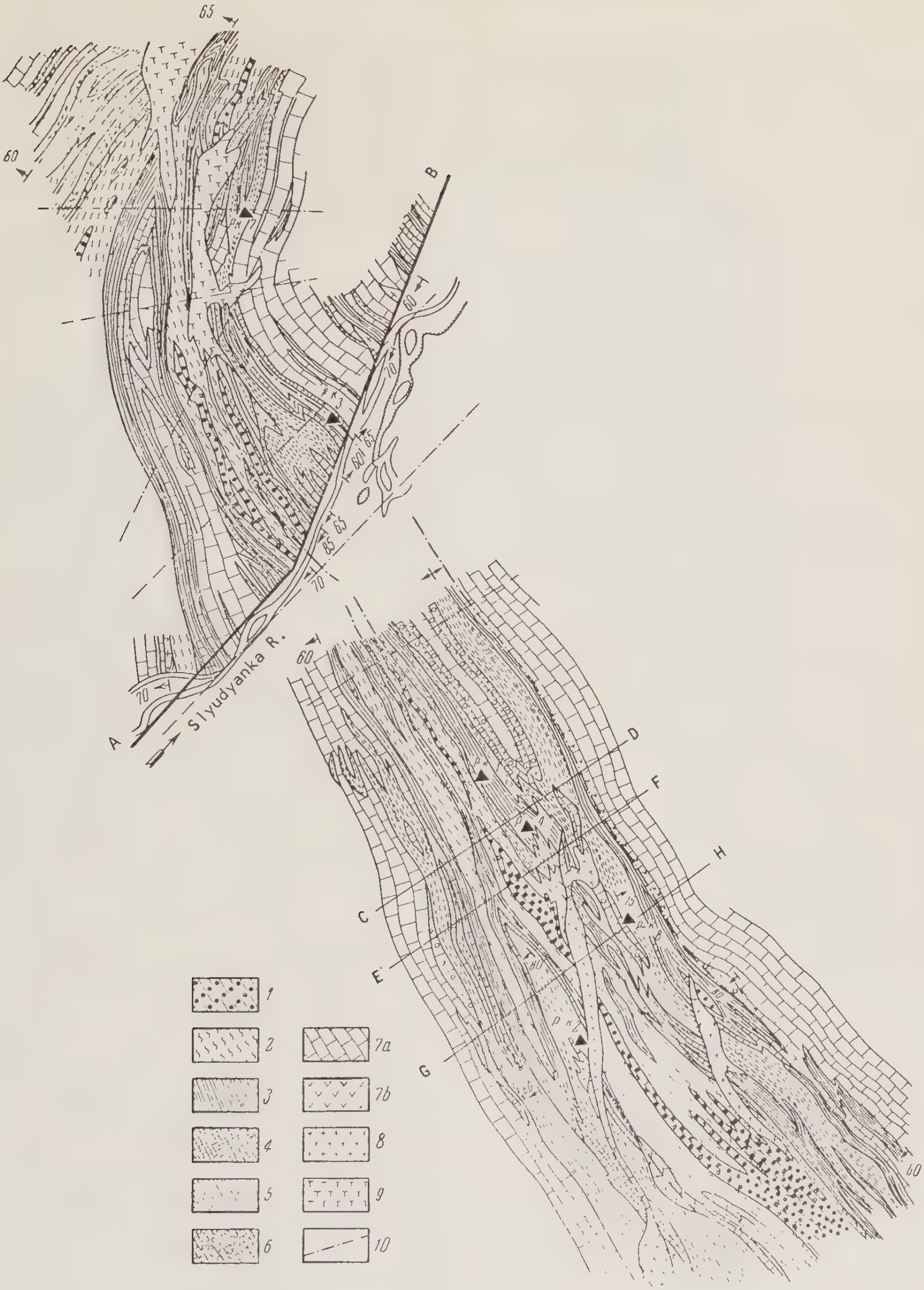

FIGURE 1. Geologic map of the ore zone of the Slyudyanka phlogopite deposit, Southern Baikal.

1-biotitic-garnet gneiss; 2-hornblende-pyroxene-biotitic coarse-flaked gneiss; 3-pyroxene-amphibole gneiss; 4-marble, overlying leucocratic biotitic gneiss turned to dolomite in the upper part; 5-leucocratic small-flaked biotitic gneiss; 6-quartzitic-diopsidic rocks; 7a-marbles underlying leucocratic biotitic gneiss; 7b-diopsidic rock, displacing marble; 8-pegmatite; 9-hybrid rock; 10-tectonic zones of fissure.

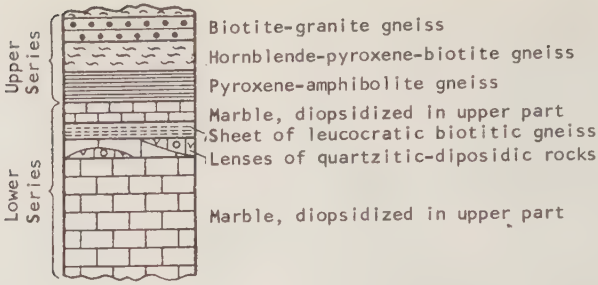

FIGURE 2. Stratigraphic core of a deposit.

marble series, closely compressed isoclinal folds composed of gneiss of the upper series predominates in the mining zone.

A supporting geologic profile (Fig. 3) constructed for the Slyudyanka valley shows that the gneissic belt of the mine field presents an alternation of three anticlinal (I, II, III) and four synclinal (1, 2, 3, 4) folds.

The first anticlinal fold, situated in the northeast part of the mining zone, is wide and open with the limbs dipping southwest and northeast. This anticlinal structure is composed primarily of small-flaked leucocratic biotitic and biotitic-hypersthenic gneiss containing local impregnations of graphite, garnet, and cordierite. In the apical part of the structure the underlying marble and quartzitic-diopsidic rocks are exposed. In the northwest on the right slope of the Pokhabikhi River the structure is closed.

The second (II) anticlinal fold is located to the southwest of the second (2) synclinal bend of pyroxene-amphibole gneiss and, in contrast to the wide and open first (I) anticlinal structure, is represented by a narrow lense elongated and closed off to the northeast and southwest. This closely compressed anticline, like the first (I) is composed of small-flaked leucocratic biotitic, biotitic-hypersthenic and biotitic-cordieritic gneiss with an outcrop in the anticlinal part of the structure of the underlying diopsidized marble layer. Its southwest limb has undergone intensive tectonic seaming with the development of highly sheared sections containing lenses of pyroxene-amphibole gneiss.

The third (III) anticlinal fold is blind for the greater part of the stretch. On the geologic map of the central part of the mining zone it corresponds to the pyroxene-amphibole belt which is the nucleus of the anticlinal fold. The underlying marble and the block of leucocratic biotitic gneiss composing the lower series are exposed at the surface in the southeast (region 2 of the mine) and in the northwest (watershed of the Slyudyanka and Pokhabikha Rivers) sections of the belt. The leucocratic biotitic gneiss in the cores of the anticlinal folds, side by side with the biotite, are sporadically impregnated with garnet, cordierite, and more rarely sillimanite.

All three anticlinal folds described above are extremely complex in structure and, depending on the depth of erosion, have various forms. The complexity of form of the folded structures is attendant on a number of causes, namely: varying thickness and plasticity of the layers and the relative position of the rocks which form these structures. Thus, the marble of the lower series, being massive and having great thickness, forms wider, more regular folds and outcrops at the surface, as shown on the geologic map and cross-sections (Figs. 3 and 4), in the core of the central apical part of the large open anticlinal folds. The thin layer of small-flaked biotitic gneiss lying in the upper part of the marble of the lower series, because of its increased plasticity, is collected in the periclinal part of the structures in closely compressed disharmonious folds which fill up the bends in the underlying marble layer, thus sharply augmenting the width of its exposure, and developing zones of undergirding.

Together with this, in limbs of closely compressed folds, small-flaked biotitic gneiss was subjected to extension and contraction in the formation of sheared tectonic zones, accompanied in a number of cases by the folding of these gneisses. The tectonic sheared zone noted earlier by A. S. Suloyev is an example of this; it was traced by us along the northeast contact of the gneissic belt in the mine field (Figs. 3, 4, 5).

Thus, we have reviewed the anticlinal folded forms that developed within the mining zone of the lower marble series. Tracing from the gneissic belt of the mining zone to the northeast and southwest (Fig. 3),

the marble forms larger anticlinal folds, complicated by secondary folding. As a result of this, repeated seams of leucocratic biotitic gneisss, and lenses of pyroxene-amphibole gneiss of the upper section are present within the large marble exposures.

Due to the plunge of the axis of the folded structure to the northwest, the marble and leucocratic biotitic gneiss of the lower series, emerging in the anticline southwest of the gneissic belt, dip in the same direction under the gneiss of the upper series and form a lock with the adjacent southwest gneissic belt. To the southeast of mine 2, as a result of the upwarp of the axis of the folded structure, the marble with the block of leucocratic small-flaked biotitic gneiss is locked with the marble and leucocratic biotitic gneiss outcropping within the gneissic felt of the mine field.

The marble which lies adjacent to the gneissic belt on the southeast is joined together into a larger anticlinal fold than the marble of the southwest belt and is complicated by smaller folds.

The upper series, represented by various gneisses -- pyroxene-amphibole, coarse-flaked, biotitic, and biotitic-garnet, form the core of the synclinal structures, developing in the belt of the mining zone four synclinal folds (1, 2, 3, 4). Two of them (1 and 2) are located in the northeast part of the mine belt, joined with the lower parts of the upper series -- pyroxene-amphibole gneiss -- and are very closely compressed. The pyroxene-amphibole gneiss forming the core of the synclinal folds are complicated by secondary folding and, as a result of the undulation of the axis of the folded structure, have an extremely variable width in outcrop.

The overlying coarse-flaked biotitic and biotitic-garnet gneiss of the upper series are crumpled, like the underlying marbles of the lower series, in the synclinal folds, together with the pyroxene-amphibole gneiss. As a result of complex folding within the pyroxene-amphibole gneiss there are sporadic appearances either of lenses of the overlying coarse-flaked biotitic and biotitic-garnet gneiss (region of Nikitinskaya mining claim), or thin seams and lenses of the underlying marble (region of mines 3 and 4).

Two synclinal folds, located in the western part of the gneissic belt of the mining zone (3, 4), in contrast to the closely compressed synclinal folds described above, appear deeper and less closely compressed. Their cores are composed of younger coarse-flaked biotitic and biotitic-garnet gneiss.

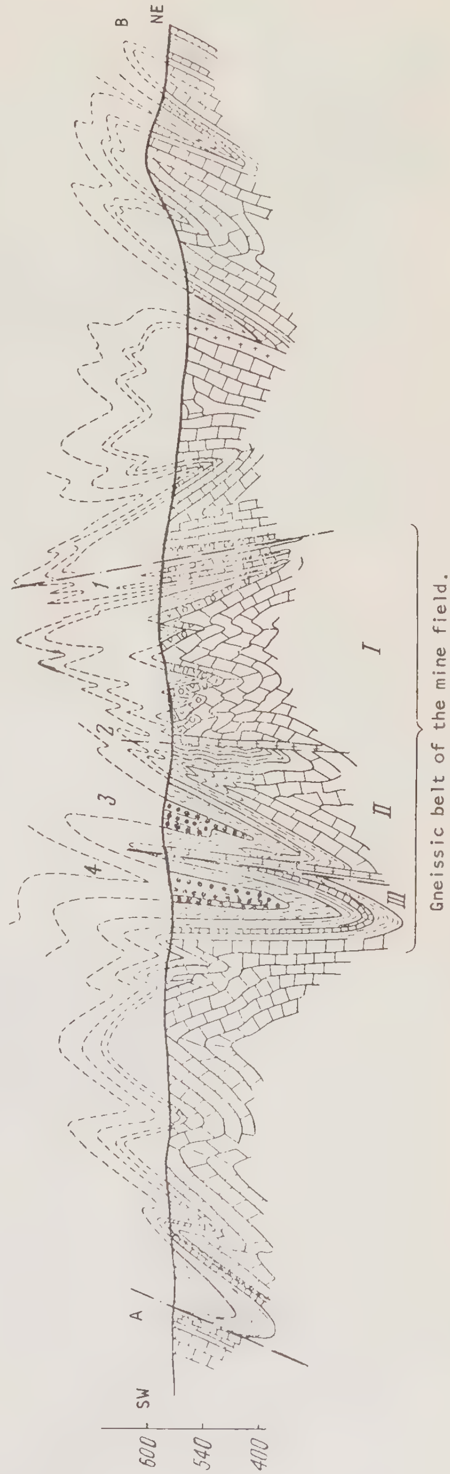

FIGURE 3. Supporting geologic profile along line AB. Conventional symbols the same as for Figs. 1 and 2; Roman numerals show anticlines, and Arabic numerals show synclines.

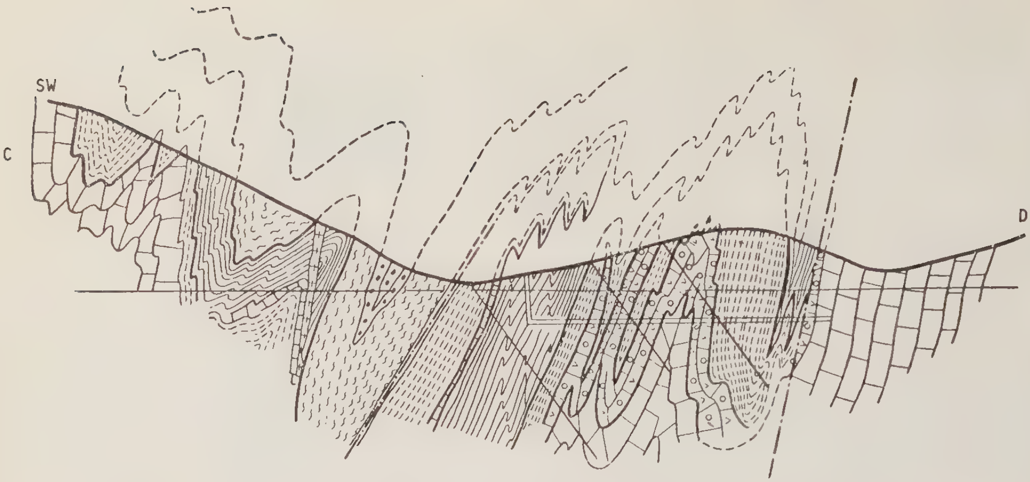

FIGURE 4. Geologic profile along line CD through shaft 4 and boreholes 1, 2, 3. Conventional symbols the same as for Figs. 1 and 2.

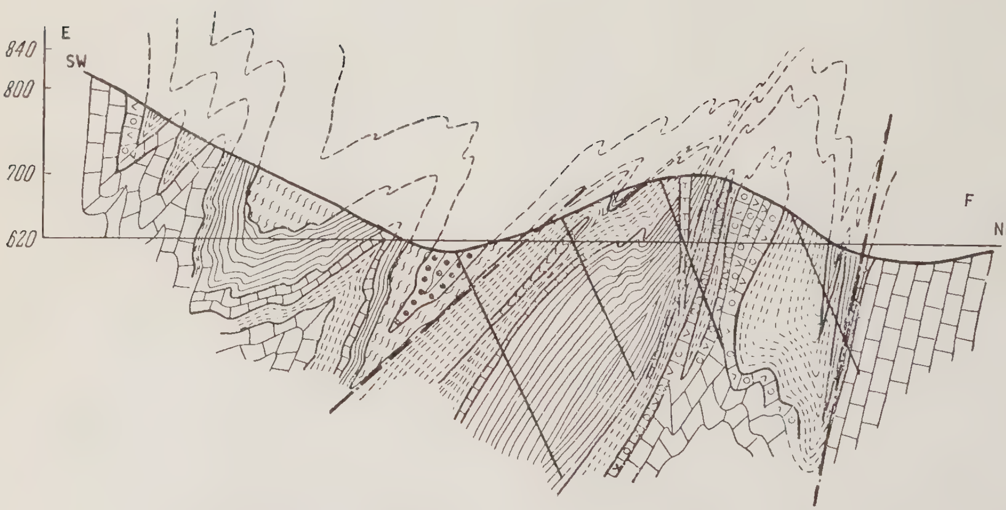

FIGURE 5. Geologic profile along line EF and boreholes 8, 9, 6, 7. Conventional symbols the same as for Figs. 1 and 2.

Depending on the undulation of the axis of the folded structure, the width of the exposure of the synclinal folds changes sharply; with a dip in the axis, wide lenses and blocks of the overlying biotitic-garnet gneiss appears within the coarse-flaked biotitic gneiss, and conversely, in an upwarp the width decreases and the biotitic garnet gneiss disappears.

Thus, the mining zone of the Slyudyanka deposit is distinguished by exceptional complexity and is characterized by the presence of numerous compressed and open folds, complicated in their turn by smaller folding. All synclinal forms in the cores of the fold are composed of gneiss of the upper series. In the cores of the anticlinal structure, rocks of the lower series, represented by marble with a seam of leucocratic biotitic gneiss, outcrops at the surface. The shape of the folds is extremely varied; folds

developing from the lower series -- marble -- are wider and, in the majority of cases, open. Folds composed of gneiss, usually compressed, are complicated by smaller folding which develops with especial intensity in the periclinal parts of the structures, where the layers forming them are collected in compressed lenses and disharmonious folds, developing peculiar semipermeable shields.

Along with the zones of undergirding of material in the periclinal parts of the structures a squeezing out of material has occurred in places of intense tectonic compression in the limbs of large folds, accompanied by a thinning of individual layers and the appearance of zones of shearing and tectonic seaming with faulting along the contact.

The cause of similar tectonic phenomena, as A.A. Sorskiy [5] points out, is the irregular plastic flow of rock materials caused by the action of tectonic compression directed normally to bedding surfaces. In fold development the material is distributed around -- it migrates from places of greatest pressure to places of lowered pressure. In this regard, the volumes of the closed parts of the folds in relation to the limbs of tightly compressed folds are greatly increased in some sections of the metamorphic part of the mining zone of the Slyudyanka deposit. This, for example, the ration of the width of a seam of pyroxene-amphibole gneiss in the region of mines 1-4 to the width of the southwest seamed limb is equal to 1:16. For the leucocratic small-flaked biotitic gneiss the width of the periclinal part of an anticlinal structure is to the width of the wing as 1:9 or 1:11.

The phenomena of tectonic seaming, compression, and shearing of strata in the limbs of large folds, on the one hand, and the development of small disharmonious folding in the periclinal parts of the structures, on the other hand, complicate in large measure the folded structure of the deposit. They are accompanied by either individual seams dropping out of the stratigraphic sequence or by a sharp increase in thickness. This is also the reason why many geologists reject suggestions of a folded structure and put forward the idea of monoclinal stratification of the rocks.

Thus, B.M. Ronenson, proceeding from a preconceived idea of cycles of sedimentation in the Archean marine basin, came to the conclusion that recurrent layers in a complex isoclinal fold are independent stratigraphic units. He therefore divided the lower marble series of the Archean of Slyudyanka into two independent sub-series: the upper, lying in the southwest part of the gneissic belt, and the lower, bordering this belt on the northeast.

The gneissic belt of the mining zone which we related to the upper series, B.M. Ronenson takes as a middle series, moreover, all similar layers recurring in a complex isoclinal fold he regards as independent stratigraphic units. Thus, the pyroxene-amphibole gneiss of the upper gneissic series, commonly crumpled into isoclinal folds, he divides into four independent layers and relates them to various stratigraphic levels (Ac^{5a}, Ac^{6a}, Ac^{7a}, Ac^{8a}). Similarly, each outcrop of pyroxene-biotitic, biotitic-garnet, and other gneiss he takes as independent layers and in this fashion he divides a great number of layers of the metamorphic series of the Slyudyanka Archean. Thus, for example, an essentially single layer of leucocratic small-flaked biotitic gneiss, emerging repeatedly in the region of mine 2 together with marble of the lower series in the anticlinal folds (Fig. 6) he takes as ten independent layers developed, in his opinion, in three cycles.

Thus, all layers of the isoclinal folding B.M. Ronenson artificially groups in "sets"; the sets he joins into cycles and in the same way he mechanically divides the complex folded stratum of the Slyudyanka Archean into a multitude of independent layers. The absence of any of the layers in the fold, caused by upwarp or sinking of the joints and squeezing of individual layers in zones of shearing, he explains as the dropping out of one or another "member of the set" from the cycle.

The data of our geological survey he applied to an artificially constructed plan of cyclic sedimentation. Without considering the phase change of the layers in direction of strike and dip, the degree of metamorphism, and the peculiarities of folding prior to reef formation, B.M. Ronenson splits up one layer into a multitude of independent layers which, in his opinion, form a monocline. All these fabrications he calls "objective criteria for analysis of strata of Precambrian rocks" [3]. These "criteria," unfortunately, find support among local geologists, but are unequivocally refuted by prospecting and operational works.

Fractures and fault dislocations

In the mining zone of the Slyudyanka deposit the following principal fracture systems have been apparent in relation to the strike of the metamorphic strata: longitudinal, transverse and diagonal.

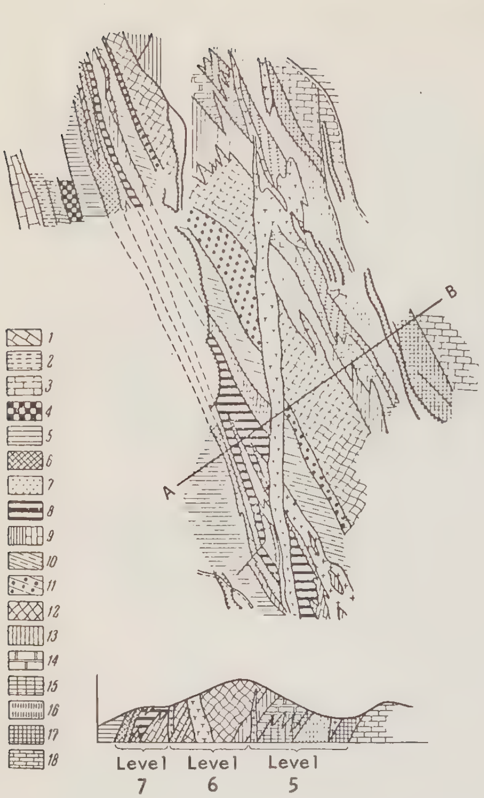

FIGURE 6. Extract from the geologic map of B.M. Ronenson, mine 2 and cross-section.

1-marble; 2-leucocratic garnet-biotitic gneiss with seams of quartzitic-diopsidic rocks; 3-quartzitic-dolomitic and quartzitic-diopsidic rocks; 4-marble with layer of quartzitic-diopsidic rocks in the upper part; 5-18 -- "sets" of various layers.

Transverse fractures, with the azimuth of the dip SE 140° and with the angle of the dip nearly vertical, are distinguished by their notable extent; they are clearly defined and are often filled with pegmatite and calcite and phlogopite veins. The presence of transverse fractures is related to the physical properties and thickness of the layer. These fractures are commonly 0.5 to 1.0 mm wide in unsheared, foliated pyroxene-amphibole gneiss. Here they are characterized by large extent and even walls. They are less clearly defined in foliated biotitic gneiss, in the form of short, winding, and sometimes branching cracks.

The longest and straightest fractures are located in areas of isoclinal folding, which in conjunction with transverse folding leads to an increase in thickness and rigidity of the layer.

Longitudinal fractures, with an azimuth of dip SW 230° to 240° and NE 60° to 70°, are also rather widely developed in the deposit; they are located in the sheared gneiss and therefore are uneven and compressed. They are found for the most part in the limbs of closely compressed folded structures and in sections of complex isoclinal folding, chiefly on the contact of the lower marble and upper gneissic series.

The longitudinal fractures, like the transverse, were important in the formation of mica, because they served as the means of circulation of postmagmatic solutions which permitted the interaction of the pyroxene-amphibole and biotitic gneiss with the underlying dolomitized marble, simultaneously with faulting of the calcite-phlogopite veins.

Diagonal fractures are found in combined systems of two fractures, with the azimuth of dip 205 and 295° respectively and the angle of dip 75 and 85°. They are less extensive than the preceding fracture systems, cleave the foliated structure at an angle of 40-50° and have even walls with smooth surfaces on which "slickensides" are commonly found.

Besides the fracture systems mentioned above, a few cracks cutting the foliated structure of the rocks at a sharp angle and dipping almost vertically occur in the crystalline belt of the mining zone.

Thrust faults and other faults stand out clearly as the major fractures in the mining zone of the Slyudyanka deposit.

Thrust faults here are quite prevalent and consist of fairly thick longitudinal tectonic zones with a dip SW 230°, ∠60° to 70° and NE 60° to 70°, ∠70° to 80°. They are located chiefly on the contact of the upper and lower series.

One thrust in the Slyudyanka deposit is traceable along the northeast contact of the gneissic belt with the marble of the lower series. It is clearly observed in the pits of mine 3, where its width reaches 18 m, and also in the underground mines of shaft 4.

The thrust described resulted from slippage between beds in the process of folding of the lower marble series along its contact with the upper gneissic series. As a result of the fold development and differential stresses on the material at the contact of the upper gneissic and lower marble series, slippage occurred that led in places to intense tectonic compression and disruption of the continuity of layers and to the formation of tectonic zones of shearing. The spaces opened by the thrust fault are in places filled with pegmatite, and igneous breccia;

in places a seaming of the layers of intensely sheared leucocratic biotitic gneiss is also present.

A second rather large tectonic disturbance of the thrust type is found in the core of the third (III) anticlinal structure on the border of the lower marble and the upper gneissic series. It has been determined in the underground and surface mines that this thrust disturbance manifests itself in the form of a wide (18 to 20 m) longitudinal fracture filled with breccia consisting of fragments of pyroxene-amphibole gneiss with a small admixture of fragments of rose-colored marble, 0.3 to 1 m wide, cemented by fine broken material and pegmatite.

Besides the large thrust faults enumerated above, a number of smaller tectonic zones are present in the seams of the linear folding in the area under study. They occur in the limbs of the folded structures and occur especially at the contact of the lower marble series with the upper gneissic series.

These cross-cutting, transverse and diagonal faults are of a second type.

Related to the cross-cutting pre-mineral fractures are meridional fractures cutting to the crystallized belt at a sharp angle. The azimuth of their dip is NE 85° and the angle of dip a steep 75° to 85°. Thick dike-like blocks of gabbroid rocks occur in them, and as a result of the migmatization of these rocks by granite and pegmatite, so-called hybrid rocks [2] have developed.

In the area under study three cross-cutting meridional faults are recorded. The first two are in the northwest part of the mining zone, on the watershed of the Slyudyanka and Pokhabikha rivers, and the third, thicker fault is traceable along the right side of the Slyudyanka River and passes along the watershed of the Nikitinsk and Uluntuy falls.

The diagonal faults are large, also pre-mineral tectonic fractures striking northeast with a dip azimuth SW 185° to 190° and NW 285° to 290° and an angle of dip, 65° to 70°. In these faults certain large phlogopite veins are located. Thus, for example, between mines 1 and 4 a diagonal fracture occurs in the midst of the pyroxene-amphibole gneiss which with smaller associated cracks appeared to be a receptacle for one of the large phlogopite veins of the deposit. The horizontal displacement of the vein is 25 m. Veins 5 and 6 are in a system of diagonal fractures at mine 2.

All the diagonal faults noted above are genetically connected with the northeast trending folds. They are located in places where linear transverse folding has occurred.

The transverse faults are both pre-mineral and post-mineral. Fractures with an azimuth of dip SE 140 and angle of dip close to vertical are related to the pre-mineral trans verse faults. These faults serve as the place of development of a number of calcite-phlogopite veins. They are clearly traced in the pits of mine 1, and in mines 4, 2, and 8.

Dislocations that cut across layers normal to their strike are related to the transverse post-mineral faults. The largest of them is the fault which passes along the valleys of the Slyudyanka and Pokhabikha Rivers. Along the watershed of these rivers there are also a number of smaller faults.

Factors contributing to development of commercial mica

The principal factors that have been favorable to mica development in the Slyudyanka phlogopite deposit have been the following: 1) chemical and lithologic composition of the rocks; 2) complex structural relationships; 3) contact of dolomitized limestone with rocks of alumino-silicate composition; and 4) tectonic fault zones. Emphasizing these conditions let us characterize the principal mineral veins of the deposit.

As has already been noted, the commercial mica occurs in definite, favorable stratigraphic layers. Such layers in the Slyudyanka deposit in the lower series are the layers of diopside rocks (which developed from dolomitic marbles) and the leucrocratic biotitic gneiss, and in the upper series -- the layer of pyroxene-amphibole gneiss. Commercial deposits of phlogopite have not been found in any other rocks within the limits of the mine field.

The largest within the limits of the pyroxene-amphibole gneiss of the upper series is the deposit of mines 1 and 4.

In the geologic cross-section through the southeast parts of mines 1 and 4 (Fig. 7) the complex structure of the belt of pyroxene-amphibole gneiss is expressed as a synclinal bend with intensive development in its core of linear and transverse folding which intersects the synclinal fold. A layer of dolomitized limestone underlies the pyroxene-amphibole gneiss, and with the latter is crumpled into folds and appears in the form of sporadic seams both in the contact zones and within the complexly folded synclinal bend of the pyroxene-amphibole gneiss. This layer is traceable in

the pits of mine 1 as a 2-meter layer in the southwest contact of the productive belt of the pyroxene-amphibole gneiss. More to the southeast, in the region of deposit 4, separate lenses are traceable, underlying dolomitized marble, both to the southwest and northeast of the contacts of the pyroxene-amphibole belt, and also within it.

gneiss, on the convex side of which the greatest number of transverse fault cracks appeared.

At mine 4 the underlying dolomitized marble layer and the pyroxene-amphibole gneiss, crumpled into complex isoclinal folds, underwent a wide mutual metasomatic

FIGURE 7. Geologic profile along line GH and boreholes 11 and 14.

Within the pyroxene-amphibole belt the marble layer is crumpled with gneiss into complex folds and forms a number of intersections of seams and lenses, filled by calcified or diopside-phlogopite rocks.

Sheared zones at the contact of the underlying dolomitized limestone with the pyroxene-amphibole gneiss served as feeders of post-magmatic solutions which circulated along these sheared zones and the multiple system of fractures adjacent to them, causing mutual metasomatic interaction between the underlying dolomitized marble and the alumino-silicate rocks. Thus, in the complex folded zones in the midst of the pyroxene-amphibole gneiss near the contact with the dolomitized limestone, blind calcite-phlogopite veins were formed.

In the light of the above account, the coordination of the calcite-phlogopite veins in mine 1 to the suspended southwest contact of pyroxene-amphibole gneiss with the dolomitized marble of the lower series becomes clear. Here the emergence of calcite-phlogopite veins was aided also, by a tectonic zone of shearing passing along the contact of marble and gneiss, and by a weak sag of the layer of pyroxene-amphibole

interaction with the participation of hydrothermal solutions enriched with magnesium. The zones of shearing intersecting the linear folding and passing along the contact of the marble and gneiss, were conducive to this interaction which finally brought about the development of the calcite-phlogopite veins in the open transverse fractures. Therefore in mine 4 calcite-phlogopite veins are spread over the whole series of pyroxene-amphibole gneiss, but are found principally in the calcareous seams.

The clearest examples of phlogopitic mica occurring in the lower series are the deposits of mines 2, 3, and 8.

The calcite-phlogopite veins of mines 2 and 8 are located in the layers of leucocratic small-flaked biotitic gneiss of the lower series and are concentrated near the contact of the gneiss with the underlying crystallized limestone. Their lower beds in the upper part have turned to dolomite where they are in contact with the leucocratic biotitic gneiss. They are represented by predominantly forsteritic calcite, some of which has been changed to light-green diopside.

In structural relations the deposit of

mines 2 and 8 presents a complex tectonic picture. Here the leucocratic small-flaked biotitic gneiss in the periclinal part of the anticlinal structures are gathered in closely compressed disharmonious folds with sharply increased thickness in the axial regions. Linear and transverse folding in combination develops complex fracture patterns.

Longitudinal shear fractures, located chiefly at the contact of small-flaked leucocratic biotitic gneiss with the underlying dolomitized limestone, served as the means of circulation of mineral-bearing solutions. The latter aided in the metasomatic interaction between the underlying dolomitized crystalline limestone and leucocratic biotitic gneiss with the development of metasomatic calcite-phlogopitic veins in transverse fractures of a fault surface. Therefore in the belt of leucocratic biotitic gneiss of mine 2 and 8 of the calcite-phlogopitic veins to tend to occur in the central part of the structure, i.e., at the contact of the leucocratic biotitic gneiss with the underlying dolomitized marble which outcrops in the core of the structure. Together with the first anticlinal structure to the southeast and northwest there is embedded a core composed of crystalline limestone; together with this layer, the phlogopite veins fade out at the surface.

In the leucocratic small-flaked biotitic gneiss ladder veins have developed which are less regular than the ladder veins in pyroxene-amphibole gneiss; this is explained by the lithologic and physical peculiarities of these rocks.

Less significant for the development of mica is a deposit which is connected with the small-flaked leucocratic biotitic gneiss of the lower series. The structure of this deposit, in contrast to those of mines 2 and 8 described above which are characterized by deep and closely compressed isoclinal folding, is represented by slightly dipping, gentle folds. These folds have also influenced the development of the mica. Thus, in mine 7, in contrast to the deposits in mines 2 and 8, whose features are interstitial ladder veins, mica occurs in the form of singular, layered lense-like beds, located at the contact of the lower marble layer with biotitic gneiss and concentrated at the axes of the anticlinal structures. According to the description of P.V. Kalinin [3], these thin lenses (up to one m) thin out even more along the strike. Some of them have the form of elongated plates, with an extent of 25 m, from which small cross-cutting veins of phlogopite branch out in places, having been formed along the system of transverse and diagonal fractures. In mine 3 the calcite-phlogopite veins are situated in diopside rock developed by means of the displacement of the lower bed of dolomitic limestone which as D.S. Korzhinskiy [4] notes, occurred partly in the magmatic stage but is principally the result of bimetasomatism.

The diopsidic rocks and calcite-phlogopite veins in the region of mine 3 lie in a limb of the large northeast-trending anticlinal structure and are related to the folds in the layers, which predetermined the development of a fan of transverse fractures in the diopside rock; these fractures were subsequently filled by the calcite-phlogopite veins.

A longitudinal shear zone in the region of mine 3, which is traceable along the contact of the pyroxene-amphibole gneiss with diopside rock, and a system of transverse and diagonal fractures were the channels of movement of pegmatite fusions causing the development of large zones of igneous breccia consisting of light-green diopside rock and pyroxene-amphibole gneiss. Subsequently, along the contacts of the pegmatite veins and the multiple system of fractures postmagmatic solutions circulated and caused a mutual metasomatic interaction between the pegmatite and diopside rock in the transverse fractures of the fault and then the formation of calcite-phlogopite veins.

Generalizing what is said above finally leads to emphasize that the principal factors in the development of commercial stocks of phlogopite are:

1. The lithologic composition of the rocks. Commercial mica is related to the layer of pyroxene-amphibole gneiss of the upper series and to the leucocratic biotitic gneiss and diopside rocks of the lower series.

2. Intersection of structural features, characterized by compound linear and transverse folds which have been subjected to intensive fracturing. The cores of closely compressed synclinal folds are related to such joints in the upper series, and in the lower series -- to cores of anticlinal structures.

3. Contact of productive layers (pyroxene-amphibole and leucocratic biotitic gneiss) with underlying dolomitic crystal limestone or diopside rock. Mica in the pyroxene-amphibole gneiss of the upper series occurs at the contact with the underlying marble seams of the lower series which outcrop within the gneissic belt in complex anticlinal folds. Mica of the leucocratic biotitic small-flaked gneiss of the lower series occurs at the contact with an underlying thick stratum of marble that has been dolomitized in its upper part.

4. Systems of pre-mineral faults, first the layered, steeply dipping thrust zones occurring at the contact of two series, and also cross-cutting (meridional), diagonal, and transverse faults. This whole system of faults and fractures has been favorable for the appearance of magmatic activity and for the circulation of mobile solutions which caused metasomatic interaction between the dolomitized marble and gneiss and, next, phlogopite veins.

The chief routes for the circulation of postmagmatic solutions were the ancient layered fracture zones. The principal mineral-bearing cavities of the phlogopite veins were the transverse fractures of the fault. Most favorable for the localization of commercial mica were the sections of complex structural intersections where isoclinal folding, a dense network of fractures, and contact of the productive gneiss with the dolomitized marble occurred.

All the genetic peculiarities of mica enumerated above, which were brought to light in the process of the geological survey, fully support the point of view of D. S. Korzhinskiy about the contactual, bimetasomatic and, in part, infiltrational origin of the calcite-phlogopite veins of Slyudyanka.

It is important to mention that, because of the idea of a complex folded structure, the perspectives of the Slyudyanka deposit broadened significantly, since in this case mica is not limited to one mining belt (the middle sub-series, according to B.M. Ronenson) but can be found in adjacent gneissic belts having favorable intersections of structural features.

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ON THE NATURE OF SMALL FOLDS IN THE MESOZOIC OF THE EASTERN TIMAN AREA

by

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This article presents arguments for the landslide origin of the folds of the Mesozoic deposits in the Eastern Timan area.

The question has been repeatedly raised in geologic literature about the nature of those small folds which are widely distributed in sedimentary rocks of the Paleozoic and Mesozoic of the Russian platform. Some authors are inclined to consider all similar folds as tectonic phenomena, others connect their formation with various surface processes -- landslides, caves, etc.

Numerous small dislocations are known, in particular, in the northeast of the Russian platform. The data of O.L. Eynor [6] and A.A. Chernov [4] on the dislocated Mesozoic in the eastern part of the Pechora depression, of A.A. Malakhov -- on the series of isoclinal folds in the upper Permian of Western Timan [1], and also of a number of investigators on the folds in the Jurassic and Cretaceous deposits of eastern Timan have produced hypotheses not only on the great mobility of this part of the platform [2, 3], but even on young fold movements within its limits [1, 3].

In this article we are concerned with the nature of some of these dislocations developed in Upper Jurassic and Cretaceous deposits of the southwestern rim of the Pechora synclinorium.

The Upper Jurassic and Lower Cretaceous deposits of the eastern Timan area are represented by sand and clay rocks of slight thickness; their section in many ways resembles the section of the Volga area.

In the continental sequence of the Middle Jurassic, represented in the lower part by dark brown clay with seams of sand and thin seams of coal, and in the upper by fine-grained micaceous sand, lies the marine Upper Jurassic. Its section begins with deposits of the Callovian -- a block of gray and greenish-gray, glauconitic sand and sandstone, in places fine-grained conglomerate, and gray clay with a thickness of 6-7 m, with Articoceras ischmae Keys in the lower part and Cadoceras in the upper part. The presence of Arcticoceras ischmae indicates that the section here begins with the very lowest parts of the Upper Jurassic.

Higher in this section lies a uniform bed of grayish and greenish, commonly glauconitic sand, then sandstone, then plastic clay. Its lower layers of 1 to 1.5 m, containing Aulacostephanus and other fossils, date to the Oxfordian and Kimeridgian. The upper part of these clays, with a thickness of 15 to 40 m, belongs to the lower Volga stage. They are covered by a layer of oil shale about 15 m thick, with seams of blue-gray and black plastic clay with numerous Cylindroteutis absoluta Fisch, Aucella mosquensis Buch. and Dorsoplanites panderi of the Lower Volga stage.

This layer is replaced by a 50-meter-thick bed of dark-gray, calcareous clay, containing in the lower part seams of shell rock, almost wholly composed of Aucella mosquensis Buch. The bases of this bed are of equivalent age to the zone of Dorsoplanites panderi, and the upper levels are analogous to the higher parts of the lower Volga stage.

Directly on the lower Volga deposits occur colluvial gray, gray-bluish, and greenish-gray clay of the Valanginian with Aucella volgensis Lah. and Polyptychites sp., which contain at the base seams of rolled nodules of phosphorite. Above them lie bluish and lilac-gray, mute clay with batches of sands, tentatively correlated with the Lower Cretaceous. The total thickness of the latter sequence reaches 20 to 30 meters in the western part of the Izhma basin and increases rapidly toward the east.

In the background of the flat-lying sand and clay rocks of the Upper Jurassic and Lower Cretaceous are a number of more or less significant dislocations. In the majority

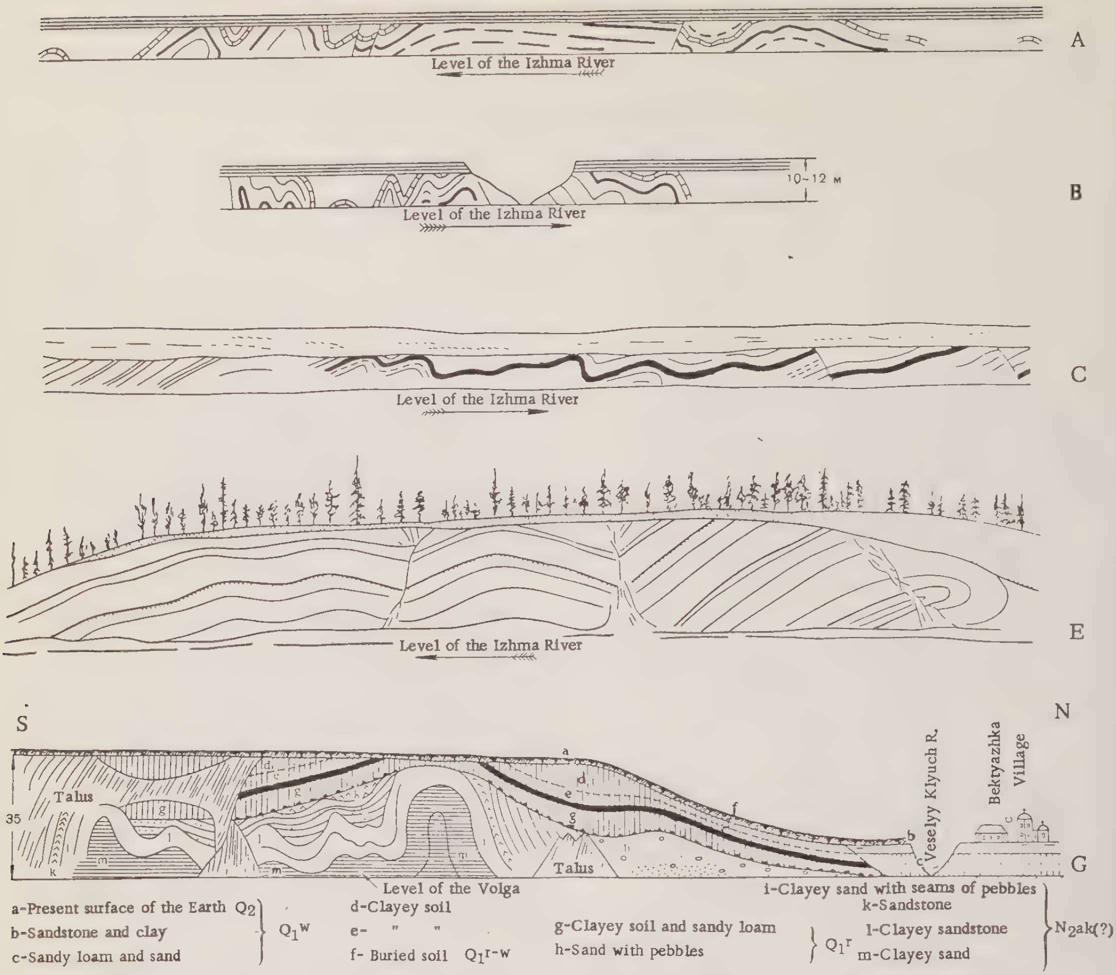

FIGURE 1. Landslide folds in the Mesozoic of the River Izhma.

A and B - below Paromes Village; C - near mouth of the Kedva (after A.P. Pavlov); E - Parus Shchel'ye (after A.P. Pavlov); E - Landslide dislocations on the River Volga, near Baktyazhka Village (after E.V. Milanovskiy; "Landslides of the Middle Volga," 1934).

of cases these are random dislocations, clearly connected with landslides -- they encroach not only on Mesozoic rocks, but also on contemporary deposits occurring in river valleys and are accompanied by sliding terraces and the appearance of woods.

Side by side with them, in the Jurassic and Cretaceous of Eastern Timan there are more regular, although smaller, folds, which have for a long time attracted the attention of investigators. They were believed to be associated with tectonic movements [3], with the pressure of moving glaciers [7], and now with landslides [2, 5].

Especially numerous similar dislocations are present in the middle course of the Izhma River, between the village of Patomes and Parus-Shchel'ye ravine; they have already been fully described by A.P. Pavlov [3]. Later this section was visited by P.S. Makeyev [2], and a few years ago the authors of this article made certain extended observations here.

In this area, about 20 km long, there are no fewer than five groups of folds. The upper, along the course, is on the right bank of the Izhma, opposite the village of Paromes, where a series of large, flat-lying sharp folds is exposed on the scarp of a terrace 12 meters above the river, under the alluvium. They are composed of black and

dark-gray very plastic, lower Volga clays, which include a seam of Aucella coquina that serves as a marking level. This outcrop along the upper part of the course, described and photographed by A.P. Pavlov [3], is almost wholly covered by landslides at the present time, whereas, in the lower part of the section there are still clearly visible ten anticlinal and synclinal folds and three small faults which complicate their limbs. Bedding measurements showed that the strike of the axes of the folds and the surfaces of the faults is fairly constant; it ranges from SW 210 to SW 240. The shape of the folds is varied: Side by side with plunging folds with rounded crests, there are sharp-pointed and box-like (folds); side by side with symmetrical folds are oblique or even overturned folds (Fig. 1-A). The width of the disturbed zones reaches 200 m.

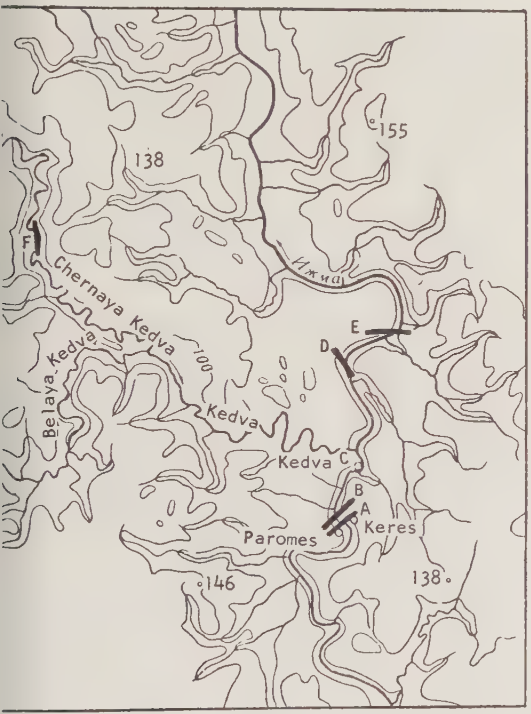

FIGURE 2. Diagram of the distribution of landslide folds in the Mesozoic sedimentary rocks of the Izhma River basin.

The second group of folds occurs within a few hundred meters, lower on the course, on the left bank of the Izhma. The width of the disturbed zone here is about 100 m, and within this distance there are nineteen folds. As the measurements taken show, the strike of their axes varies from SW 200 to SW 230. Their form here is varied, but compressed, tight folds predominate, with the dips of the limbs sometimes vertical, but the upper fold on the course is actually thrown back to the northwest (Fig. 1-B).

Farther down the river there are three groups of folds, analogous to those just considered. The uppermost of them, described by A.P. Pavlov, is located near the town of Ust'-Kedva: "Between the upper and lower Kedva rivers, the Izhma at first cuts a series of layers dipping away from the river at approximately 40° to 45°, farther on the dip is less, and a number of irregular folds or flexures occur (up to 6), still farther on there are two faults" [3]. From the sketch of A.P. Pavlov (Fig. 1-C) a great similarity is apparent between these folds and the folds of the right bank of the Izhma opposite the village of Paromes. Unfortunately, A.P. Pavlov does not indicate the strike of the axes of the folds at this outcrop, which is, at the present time, completely covered by rock waste.

The fourth group of folds is below the mouth of the branch Pukem-Iol', on the left bank of the Izhma. Here a row of plunging folds striking north-northwest are made up of "sand-clay and phosphoritic layers" of upper Jurassic and lower Neocomian [3].

Finally, another 10 km lower on the course on the right bank of the river is the large outcrop Parus-Shchel'ye, its length is a few hundreds of meters, its height about 30 m. According to the data of F.N. Chernyshev, A.P. Pavlov, and P.S. Makeyev, who described this outcrop, sandstones of the Cretaceous which make up Parus-Shchel'ye are crumpled into three folds having a latitudinal strike (Fig. 1-E). The upper fold along the course is recumbent to the south [2, 3].

Comparing all the enumerated groups of folds, it can be seen that all of them, undoubtedly, are of one type and are similar. It is notable that the strike of the folds within each group is relatively constant or varies within small limits, but is sharply different for the various groups. Moreover, only the folds exposed near the mouth of the Pukem-Iol' have a north-northwest strike, close to the strike of the Timan dislocations. The folds exposed near Paromes are almost perpendicular to the Timan trend, and the latitudinal folds of Parus-Shchel'ye trend at a considerable angle to it.

The strike of the various groups agrees approximately with the strike of large sections of the Izhma valley in which they are located (Fig. 2); it thus agrees basically with the direction of the main bank, which, near Paromes nearly approaches the right bank of the Izhma and has a height of 48 m [2].

Turning now to the question of the origin of the Izhma folds, let us remember that their relatively regular form and the significantly large area encompassed by each group are the chief arguments of the proponents of the tectonic nature of these outcrops [3]. Nevertheless, groups of folds very similar to these have been observed in contemporary landslide masses, for example, among the large landslides of the Volga area.

Thus, neither the form nor the dimension of the Izhma outcrops are unusual for landslide dislocations, and their undoubted connection with geomorphologic and not tectonic elements favors the idea of the landslide origin of these outcrops, as expressed once by F.N. Chernyshev, and later by P.S. Makeyev. The development of the landslide folds of Izhma which are covered with undisturbed alluvial deposits of the 12-meter river terrace, is not connected with contemporary, but with more ancient stages of the valley's history.

The dislocations of the Mesozoic deposits of the Izhma, therefore, do not present a sufficient basis for considering that the East Timan area was the scene of any intensive tectonic movements in late Mesozoic time.

It is possible to suppose that many other small folds of the platform are connected with tectonic movements only insofar as the latter have an influence on the surface processes which produced the dislocations we have been discussing.

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CLASSIFICATION AND NOMENCLATURE OF THE MICROCOMPONENTS OF COAL

by

A. I. Ginzburg

Recently, in the literature of coal petrography some suggestions have appeared concerning the classification and terminology of the microcomponents of coal. These suggestions were evoked by the great variety of coals, the existence of a considerable number of coal petrography laboratories in geologic, coal chemistry, and enrichment institutes, and coke-by-product works. The small number of publications on the coal petrography and the lack of sufficient communication and exchange of experience among coal petrographers has also contributed to this.

Nevertheless, the parallel existence of several classifications and nomenclatures can lead both petrographers and researchers in allied fields to baffling questions and, what is more dangerous, to confusion in specific studies.

With this in mind, we have taken the liberty of checking all published classifications and nomenclatures for the purpose of bringing about their agreement and to clarify in which type of studies the use of any particular classification is possible and necessary.

For many years the majority of coal petrographers have divided the microcomponents of humous coals into four basic groups [3, 4].

Three of them represented the products of the conversion of plant textures: gelatinization, weak and strong fusitization with preliminary gelatinization, and in one group spores, cuticles, and resinous corpuscles are combined. Subrounded bodies and sclerotics are divided independently. Thus, groups of microcomponents bear the following names: gelatinization having a red color under transmitted light, and gray under reflected light; weakly fusitized -- correspondingly brown and white of various shades; fusitized -- black and intensely white, and cutinized -- yellow and dark gray.

The criteria which permit making such a division into groups are color, relief, and outline of the microcomponents under the microscope under both transmitted and reflected light.

Further subdivision within the groups, as a rule, has been made according to the degree of preservation of the plant structure of the altered and converted parts of plants. The remains of textures with clearly visible plant structure related both to the group of gelatinized and weakly fusitized are called xylin, and those with indistinct plant structure xylovitrain. Strongly fusitized plant textures with distinct plant structure are designated by all researchers as fusite and xylenofusite, and with indistinct cellular structure -- xylovitrain-fusite. Plant textures which preserve the contours, although with almost invisible structure, are called vitrain if they are gelatinizations and vitrain-fusite if they show subsequent strong fusitization.

Microcomponents which have come from plant textures but have not preserved the plant structure and play the role of cement are called basic mass. If they are analogous in color and relief to gelatinized plant textures, it is usually termed a gelatinized or weakly fusitized basic mass. In this same way, i.e., by color, relief, and degree of preservation of structure, subrounded bodies, which by many researchers are very tentatively called sclerotics, are distributed according to groups.

A new terminology for the microcomponents of coals was presented at the Second Coal Geology Convention in 1955 by I.E. Val'ts.

In this classification I.E. Val'ts considers: 1) source material; 2) processes of conversion; 3) degree of preservation of plant structure. She introduces a number of new terms which point to the botanical or anatomical nature of the relic structures.

I.E. Val'ts proposes to adopt the name

of lignite for yellow stalk-textures; to the products of the gelatinized parenchymatous textures is given the name parenchymates, in distinction to xylinites and vitrainites, developed as a result of disintegration of predominantly stalklike textures (wood pulps, periderms, and others). For the weakly fusitized microcomponents, I.E. Val'ts reserves the names of long standing in foreign literature -- semifusites and fusites.

Further subdivision of clearly-distinguished fragments of textures within each of the enumerated groups was made by consideration of the degree of preservation of their original cellular texture. Four types of structure were distinguished and denoted by the first four letters of the Greek alphabet: α - structure -- well-preserved plant textures in which the cavities of the cells are empty (they were formerly designated as xylenitic); β - structure -- plant textures with indistinct preservation of the plant structure (xylovitrain); γ - structure -- good preservation of cell structure, cell cavities filled with amorphous organic substance (structural vitrain), - structure -- similar formations having contours (vitrain). I.E. Val'ts proposes to call the aggregate of all the enumerated structures the attrite structure.

In the basic mass the name desmite, proposed by I.I. Ammosov [1], has been adopted for materials playing the role of a connective substance.

All subrounded structures are assigned by I.E. Val'ts to an independent group and, as in foreign works, called sclerotics, but with the addition of a determination of their color in transmitted light, and determination of the degree of preservation of structure (α, β, Δ). I.E. Val'ts divides spores and pollen (exinite) into independent groups: cuticles (kutinite), suberinized bark tissues (suberinite), resinous bodies (resinite).

The classification and nomenclature for the microcomponents of coal proposed by I.E. Val'ts certainly represents a step forward in the study of coal. The great merit of this nomenclature is the brief description of the degree of structure preservation without dependence on the type of substance (α-, β-, γ-, Δ- structures) and the separation into independent groups of lignitic and parenchymatous textures.

From our point of view, the addition of the subrounded bodies, called sclerotics by the author, to the groups of basic substances is expedient. It is better to separate those typical sclerotics not subject to question, as an independent group.

However, the categorical division of all textures into lignitic and parenchymatous is open to question.

Study of the coal deposits of various basins has shown that it is not always possible to distinguish safely the botanical or anatomical affiliation of the microcomponents. In such cases, when it is not known whether the microcomponents belong to the parenchymates or the stalks there is no appropriate name for the microcomponent in the proposed nomenclature. This refers to gelatinized microcomponents and in even greater degree to the fusitized. Among the fusitized, unfortunately, we do not generally differentiate the original plant material at the present time.

Ignorance in determining the original parts of plants among the converted remains upsets the orderliness of the proposed nomenclature and classification; because of this it is impossible to use them.

However, it is quite certain that in those cases where the coals are relatively little changed by the initial processes of conversion, and also relatively little metamorphosed (brown coals and coals up to PZh) the classification of the microcomponents of I.E. Val'ts is quite convenient, permitting a deeper study of the substance of coals.

Where the botanical affiliation of the remains of plant tissues in a group of gelatinizations is unknown, the name "gelite" should be applied to them. In considering the preservation of the plant structure it would be lettered thus: α - gelite, β - gelite, γ - gelite, Δ - gelite.

In May 1956 at the All-Union Conference of Coal Petrographers at the Institute of Fuel Minerals, Academy of Sciences, USSR, a nomenclature was worked out for the petrographic components of coal for technological purposes (Table 1). Five groups of coal microcomponents are recognized, varying according to the properties of the substance.

The group of vitrains includes the microcomponents that pass into a plastic state through thermal action. To it are related all structural (tellinite) and non-structural (kollinite) gelatinized residues.

The group of semi-vitrains comprises microcomponents which on heating do not pass into a plastic state, but manifest a tendency to cake. This includes the very weakly fusitized structures; under a microscope the microcomponents of this group, both in color and in relief approach nearer to the gelatinates, although they also have intermediate properties.

Table 1

Comparison of the classification and nomenclature of coal microcomponents

Yu. A. Zhemchuzhnikov, 1948; A.I. Ginzburg, 1951		I.E. Val'ts, 1956		All-Union Conference of Coal Petrographers 1956		Terminology proposed by International Committee for Coal Petrology 1955	
Group	Name of microcomponent	Group	Name of microcomponent	Group	Name of microcomponent	Group	Name of microcomponent
Gelatinized	xylin	Products of gelatinization	α - lignitite α - xylinite α - parenchyte	Vitrainite	telinite	Vitrainite	telinite
	xylovitrain		β - lignitite β - xylinite β - parenchyte				
	structural vitrain		γ - lignitite γ - vitrainite γ - parenchyte				
	non-structural vitrain		Δ - lignitite Δ - vitrainite Δ - parenchyte		kollinite		kollinite
	Basic mass } xylo-vitrain, uniform		ligno-, vitro-, xylo-, parencho- } attrite				
	subrounded bodies		α -, β -, Δ - } rubro-sclero-tinites[1]				
Weakly fusitized	xylin	Products of gelatinization and consequences: Weakly fusitized	α - semifusite	Semi-vitrainite[2]	semi-telinite	Inertite	semifusite
	xylovitrain		β - semifusite				
	vitrain		γ -, Δ - } semifusite		semi-kollinite		
	subrounded bodies		α -, Δ - } orthosclerotic				
	basic mass		semifusiteattrite semifusitedesmite		semi-kollinite		
Fusitized	fusite	Products of gelatinization and consequences: Strongly fusitized	α - fusite	Fusite	fusite and semifusite		
	xylofusite						
	xylovitrainofusite		β - fusite				
	vitrainofusite		γ -, Δ - } fusite				fusite
	subrounded bodies		α -, β -, Δ - } nigro-sclerotinite		sclerotinite		sclerotinite
	basic mass		fusite-attrite fusite-desmite		micrinite		micrinite
Cutinized	spores	Lipoid	exinite	Leiptinite	sporinite	Exinite	sporinite
	pollen				polinite		
	cuticle		kutinite		kutinite		kutinite
	resinous bodies		resinite		resinite		resinite
	suberinized substances		α - suberinite Δ - suberin-attrite		suberinite		
Other	sclerotics						

All the sclerotics of I. E. Val'ts are related to other microcomponents.

Very weakly fusitized microcomponents. This group of microcomponents
y partly corresponds in content to the weakly fusitized group.

The fusitite group consists of microcomponents which, during heating, do not pass into a plastic state. It includes the weakly and strongly fusitized microcomponents.

The leiptinite group consists of the coatings of spores, cuticles, suberin scraps, and resinous bodies.

The alginic group includes seaweeds and the nonstructural sapropelic basic mass.

This classification of the microcomponents of coal, as is apparent from the preceding breakdown, is of restricted utilitarian significance and can be used only for solving technological problems, chiefly questions of coking.

For the study of coal substances, the solving of genetic questions, and also for comparing coal layers, it is not adequate to divide the microcomponents of coal into groups with almost no regard for the individual components. For example, all weakly fusitized tissues are semifusites combined with fusites, but they require independent division. Thus, the use of this classification and nomenclature in solving a number of questions both in coal geology in general, and in coal petrography in particular, is impossible; this could be a step backwards.

Moreover, the proposed designations, from our point of view, are unfortunate. Vitrainite comes from the word vitrain and, consequently, all microcomponents of this group should come from vitrain; this is not what actually happens. Vitrain represents a plant texture at a specific stage of transformation. It possesses quite definite chemical and physical properties, often not compatible with the properties of other microcomponents of this same group, in particular xylin, xylovitrain, and the basic mass. It would be more appropriate to combine all microcomponents of this group under a name characterizing the total process of change of plant substance, which is the process of gelatinization.

We still have to take up the terminology presented by the International Committee of Coal Petrologists in 1955 and approved in London in 1956.

The Committee separated three basic groups: vitrainite, inertite, and exinite. The inertite group consolidates all the fusitized microcomponents. Microcomponents within the group are separated according to structural principle, i. e., by degree of decomposition,. only in the vitrainite group, just as in the nomenclature described above, adopted at the conference of coal petrographers in 1956.

The international classification is even more narrow than that adopted in 1951 at the conference of the Institute of Fuel Minerals. The division of all lignite-cellulose microcomponents of coal into two groups with the designation of all fusitized microcomponents as inertite testifies that coals are regarded exclusively from the point of view of their suitability for coking. There is absolutely no consideration of the wide possibility for using coal in various branches of industry. And although the fusitized microcomponents are inert in coking, they can be very active in other fields of use, in particular in the power industry or as a reducing additive, etc. Therefore, the application of the name "inertites" to the fusitized microcomponents seems inappropriate. The proposed consolidation of microcomponents into groups also seems inadequate.

This exhausts the published list introducing independent classifications on the nomenclature of microcomponents of coal.

All the proposed classifications, depending on the purposes for the study of coal, justify their existence. Each of them suffers from some insufficiency; some have too little detail in the separation of microcomponents, others have useless detail. The classification of I.E. Val'ts, the most detailed, very often cannot be used because, at the present level of knowledge, it is not always possible to determine the original material.

The common positive feature of all the classifications of coal microcomponents is the bringing of the gelatinized microcomponents into one group. On this question there is no disagreement.

It should be kept in mind that the use of coal petrography in various fields of scientific knowledge does not demand the same degree of detailed study. It is necessary to analyze, in setting up any industrial plant, which microcomponents should be discriminated and in what circumstances the study is to be used.

We have made an attempt to show what the degree of detail of a study should be and to what degree between microcomponents are valuable in various investigations. For this is proposed an old, but refined, classification of the microcomponents of coal (Table 2). For weakly fusitized microcomponents the prefix semi is introduced, indicating the weak fusitization of the substance of the microcomponent (semixylofusite, semixylovitrainofusite, semivitrainofusite). The designations of gelatinized and fusitized microcomponents is unchanged. Spores and pollen, cuticles and resinous bodies are divided into independent groups (Table 2). Suberin

icrocomponents should be referred to the ɔore-cuticle group. When typical sclerotics ·e present they must be divided individually.

not wise. Such simplified studies of coal petrography do not reveal the broad distribution of microcomponents in coals, the condi-

Table 2

Microcomponents		For questions of the origin of coal and comparison of coal beds		For the purposes of the practical use of coal		
		In lumps		Briquets	Lumps	
		micro-sections	uncut sections	micro-uncut sections	micro-sections	uncut sections
Gelatinized	xylin xylovitrain vitrain-structural vitrain-non-structural subrounded bodies basic mass	divide separately all structural microcomponents and the basic mass	divide only microcomponents	consolidate microcomponents	divide separately all structured microcomponents and the basic mass	consolidate microcomponents, especially in coals of high degree of development
Weakly fusitized	semixylofusite semixylo-vitrainofusite semivitraino-fusite subrounded bodies basic mass	divide separately structural and non-structural		consolidate microcomponents		
Fusitized	fusite xylofusite xylovitraino-fusite vitrainofusite subrounded bodies basic mass	divide separately fusite with structural microcomponents from non-structural		separate fusite; combine other microcomponents		
Spore-cuticle	pollen spores cuticles suberin	separate spores, cuticles, and suberin substance		combine microcomponents		
Resinous	resinous bodies resinoid formations	separate resinous corpuscles		combine microcomponents		

Before proceeding to concrete recommen-·tions, it is necessary to underline espe-ally that the study of coals in briquets has gnificance, for the most part, only for ·rrow utilitarian purposes. To reduce coal ·trography -- the science of the nature of ·al substances and their origin -- to formal ·gistration of the various groups of micro-·mponents, possibly only in briquets, is

tions of accumulations of their parent substance, or the processes of their subsequent change. If all coal petrography is reduced to the study of briquets and tabulation of the groups of microcomponents this means the reduced capability of this science in solving other very important and practical questions.

For geologic questions connected with the

origin of coal beds, comparison of them, metamorphism, etc., study of coal should be conducted in lumps and not in briquets.

In studying coals in microsection it is more correct to make a division of all structural and non-structural microcomponents forming coals; they should rarely be in the same groups. This is of special importance in considering gelatinized, cuticle, and resinous microcomponents; no less important is the separation of real fusite and non-structural fusitized substances, designated by many investigators as the basic mass.

There is a smaller possibility of solving these questions in uncut sections and often only certain microcomponents are distinguished, namely: real fusite stands independently, all fusitites in one group, gelatinates in another, and cuticles in the fourth. However, it is more expedient, in studying coal of a relatively low degree of development in uncut sections, to divide the microcomponents within separate groups with the help of oil immersion. In studying coals of a high degree of development in uncut sections, to avoid mistakes by combining gelatinized and weakly fusitized microcomponents into groups.

In applying coal petrography to the solution of questions of the practical use of coals connected with coking and semi-coking, if the study of coals is made in uncut sections from lumps and in briquets it is more expedient to separate the groups of microcomponents almost without regard to the individual microcomponents, namely: gelatinized, weakly fusitized, fusitized, cutinized, and resinous (Table 2). Only in these cases is it possible to use the designations proposed at the conference of coal petrographers in 1956.

In studying coals for these same purposes but in microsections, it is necessary to distinguish the basic mass with xylovitrainite structure from real xylovitrain within the group of gelatinized microcomponents, i.e., microcomponents which affect the coking process in a special way. Moreover, in microsections and in uncut sections from briquets it is expedient to distinguish real fusite. To avoid errors it is better to consider the remaining microcomponents only by groups.

In conclusion it should be emphasized that none of the proposed systems of classification is dying out. However, if the classification adopted by the coal petrographers in 1956 can be used for coals in solving questions of coking, I.E. Val'ts classification chiefly for lightly metamorphosed brown coals, studied exclusively in transmitted light and in very thin microsections, then the old system of classification for microcomponents of coals, only slightly modified, remains, at the present time, of most applicability. It can be used in solving various questions of geology and technology independent of the method being applied in research, although it suffers from a few complicated designations.

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BRIEF COMMUNICATIONS

ON THE METHOD OF OBTAINING MONOMINERAL FRACTIONS FOR DETERMINING THE ABSOLUTE AGE OF ROCKS BY THE ARGON METHOD

by

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Recently, it has become apparent that of ll the methods in use for determining the bsolute age of rocks the most advantageous or wide use in geologic practice is the rgon method.

Numerous experiments in determining bsolute age by this method, both of various ninerals of magmatic origin and also of all amples of magmatic rocks, show that by no neans all potassium-bearing minerals can e used for this purpose with equal success nd that the most suitable in this respect re the micas -- especially muscovite and iotite [1, 4].

The potash feldspars, as a rule, give ges that are too low, as is apparent from omparison of the age of potash feldspars nd micas taken from the same piece of re [5], especially in cases where their ariable geologic growth rate must be conidered exceptional (in particular, in pegmatites).

The reasons for the loss of radiogenic rgon by potash feldspars, which causes the owered age figures, cannot presently be xplained, but from all the circumstances it s quite probable that they are connected ith the phenomena of reconstruction with he passage of time, and are dependent on he thermodynamic conditions of the crystalized framework of potash feldspars.

Potash-bearing volcanic glass of igneous ocks, as our studies have shown, also oses radiogenic argon during the course of me.

The natural inference from these and a number of other factual data, on which there is no room to dwell here, is the conclusion [4] that determination of absolute age by the argon method does not give us the true age of the rocks but the value of its maximum age limit, which in many cases can be significantly less than its true age.

To obtain more reliable and more comparable age figures, it is necessary to make the measurements on monomineral fractions and, primarily, on the micas.

Proceeding from the fact that micas are stable retainers of argon, we expressed the hypothesis in 1955 that glauconite, in structural aspects extremely close to micas, may retain argon with sufficient stability and may be the key in dating sedimentary rocks by the argon method [4]. By our experiments, and also by the results of measurements made by other investigators, this hypothesis was eventually confirmed [6]. Although the wide use of glauconite in determining the absolute age of sedimentary rocks demands broad methodical investigations, in particular of the comparison of figures obtained for glauconite with the figures obtained for micas of the same age, the promise of glauconite in this respect is unquestioned.

From the exposition above it is clear that for transitions to large-scale determination of the absolute age of magmatic rocks and for making methodical investigations through the use of glauconite for dating sedimentary rocks by the argon method, it is necessary to guarantee the possibility of obtaining monomineral fractions of muscovite,

biotite, and glauconite by the least laborious means and in sufficient quantity.

In this connection there arose the problem of obtaining sufficiently clean fractions of the minerals named from crushed specimens of rock, taking into account, above all, the predominant grain size, because fractions that are too small present difficulty both in control of cleanliness of the separator and in preparation of the material for melting to separate the radiogenic argon. In this respect the most suitable fractions proved to be 1.0 to 0.5 and 0.5 to 0.25 mm.

The possibilities of selecting a means of enriching the fractions were limited by the fact that centrifuging in gravity solutions seems undesirable in many respects, especially for glauconite which has a high capability for cation exchange.

Inasmuch as both biotite and glauconite belong to the group of weakly magnetic minerals, this property was taken as the starting point in devising instruments for obtaining the maximum clean fractions of these minerals. A simple method of dry separation in laboratory conditions was chosen.

In the separator we constructed (see Fig. 1) the magnetic pole is made of two flat terminals serving as the base for two steel tubes placed at a distance of 1.3 mm from each other. Because of the small clearance in it a strong heterogeneous field is created mainly outside the generating poles. The partial discharge of the magnetic lines of force beyond the limits of clearance of the polar tubes accomplishes the charging of the inner cavity and clearance by the diamagnet. For this purpose, the polar tubes are fitted to a bronze axle with a belt. An endless belt is put through the poles and is drawn up tight to turn the spindle.

The feeder is on the side; rock from the bin is fed by means of a measuring hopper to the belt in the section where the latter is crossed by the magnetic field; by this a magnetic fragment is held to the belt and is carried to the receiver while nonmagnetic fragments are poured into a second receiver.

Omitting details of the rating of the magnetic field intensity [2] necessary, let us mention that this rating indicates that for separating of the most lightly magnetized particles, the plan we adopted requires a magnetic field intensity in the gap of the order of 12-13 thousand oersteds. With this value of field intensity and the area of interaction of the poles equal to 0.5 cm^2 from the coil of the electromagnet, the creation of a magnetizing force of 5,000 ampere turns is required, considering the loss in the magnetic circuit.

The coil is constructed in two sections connected in parallel to a rectifier with an intensity of 12-24 volts. Thus, by connecting one or both sections and also by varying the amount of current in the coil by means of a rheostat, it is possible to obtain various values of magnetic force -- from a minimum up to 5,000 ampere turns.

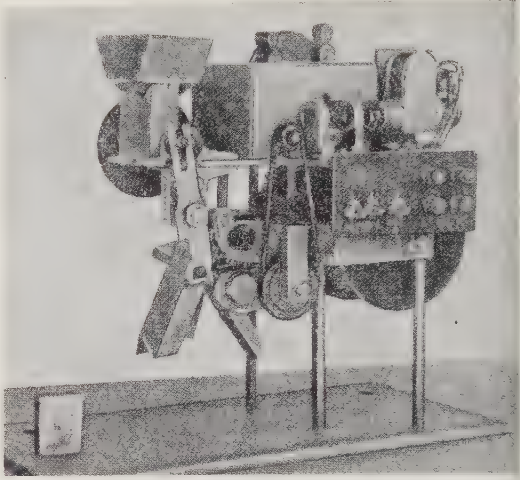

FIGURE 1. Separator for obtaining the maximum quantity of clean fragments of biotite and glauconite.

The kinemetics of the instrument is fairly simple, and inasmuch as it was developed with consideration of the parts and materials on hand, it can be simplified further.

As is apparent from the kinematic diagram (Fig. 2), the drive of the separator comes from a miniature motor, 1, with a fairly high speed. To reduce the speed and increase the torque, a worm couple, 3, is attached; the worm is connected directly to the axle of the motor through an elastic clutch. The worm gear of the couple is fitted to the axle of a conical drive spindle, 4. This spindle transmits the movement of the sprocket wheel of the measuring hopper, 9, through an intermediate roller, 18; conical spindle, 5; twin pulleys, 6 and 7; belt drive, 23, and pulley, 8. The speed of the sprocket wheel is selected depending on the actual conditions of separation for the given material. Regulation of the speed of the sprocket wheel is effected on the one hand by shifting the intermediate roller 18 along the generating coupling cones, and on the other -- by transferring belt drive 23 from pulley 7 to pulley 6.

Moreover, the conical drive spindle 4 transmits the movement through intermediate roller 18, cone spindle 17, pulley 16, belt

4, pulley 15 and the twin cylindrical cones 4 to drive spindle 19. Over this bevelled pindle and the cylindrical terminals of the lectromagnet an endless belt 13 is thrown, vhich moves in the direction indicated by an rrow. Its speed may also be regulated by hifting the corresponding roller 18.

In separating biotite with the separator, n the electromagnetic fraction along with he biotite may be magnetite, pyroxene, and ornblende, and also concretions of biotite vith other minerals, in particular quartz nd feldspar. Preliminary separation of the ighly magnetized fraction is made in the ame separator with the coils switched off residual magnetism is used) or with a low alue of magnetizing force. Partial riddance f pyroxene and hornblende is also effected y regulating the value of the magnetizing orce. However, complete riddance of the oncretions is practically impossible by this nethod.

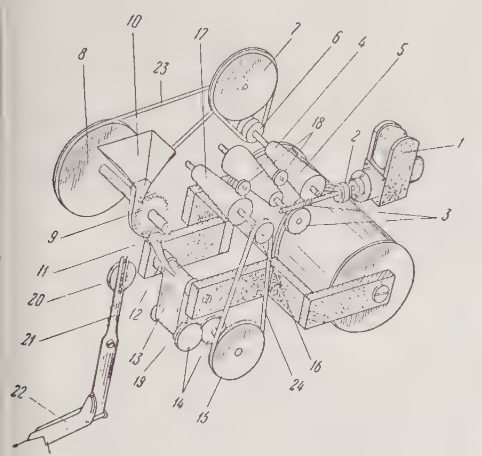

FIGURE 2. Kinematic diagram of a separator for obtaining clean mineral fragments.

Consequently, a method of separating ninerals according to their shape is applied or which a sieve is used, made of a ibrating chute, the bottom of which consists f thin plates set in the shape of a tile with arrow cracks in between. Inasmuch as the oncretions, as a rule, have isometric rains, they do not go through the cracks ut roll along the bottom of the sieve, while ne flakes of biotite, of small thickness, liding along the sloping surface easily penerate the cracks, thus being separated from ne concretions and also from the isometric rains of pyroxene and hornblende still nseparated from biotite. The use of this imple device greatly increases the purity f separating of the biotitic fraction.

The sieve is a structural part of the separator; by throwing the belt over from pulley 15 to pulley 20, a rocker, 21, is set in motion to which the chute is directly fastened. The fraction being separated drops through a measuring hopper along a removable tray not shown in the diagram (interchangeable with tray II) to the sieve hopper.

The simplicity of construction of the separator makes it easy to produce in laboratory conditions. With an average production of 2 kg of separating material an hour the separator used 0.15 kwt drawing from a VSA-6m rectifier.

Rather extensive use of this separator has demonstrated its complete reliability in operation, ease of construction, and the possibility of rapid selection of optimal conditions of separation of the various samples. We are successful in obtaining a 95-98 percent fraction of biotite and practically a 100 percent fraction of glauconite. However, the fractionation of muscovite cannot be guaranteed by this method, with the exception of the sieve which allows, nevertheless, a considerable enrichment of a muscovite sample, obtaining in some cases even up to 50 percent of its concentration. Therefore we prepared a second instrument, an electrostatic separator, extremely simple in construction (Fig. 3).

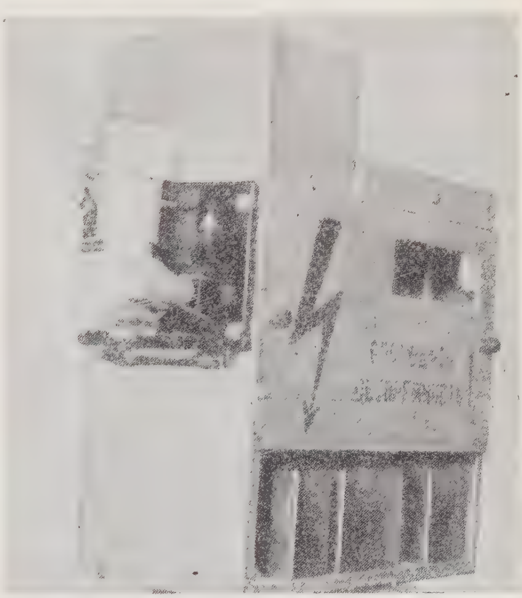

FIGURE 3. Electrostatic separator for obtaining muscovite fractions.

As is known, the operation of an electrostatic separator results in obtaining the

trajectories of charged particles moving in an electrostatic, heterogeneous field of high intensity.

The recharging of the particles of the material being separated is accomplished by means of their direct contact with the charging metallic electrode of the separator, where they receive a uniform charge of an amount depending on the conductivity of the particles, and for the non-conductors, on the size of the surface of their contact with the electrode.

Without describing in more detail, the theory of this process which is considerably more complex than in the case of the magnetic separator, let us turn to the description of the construction of the electrostatic separator which we built.

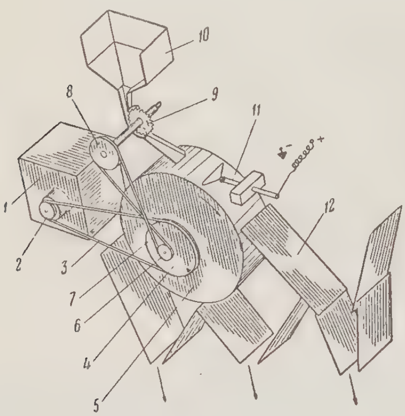

FIGURE 4. Diagram of an electrostatic separator for obtaining muscovite fractions.

As may be seen from the diagram (Fig. 4), the separator consists of a metallic drum 5, turned by a belt drive 3 from a motor with a reducer 1. The speed of the drum is low, in the order of 20-30 rpm, and is regulated by a rheostat connected to the feed circuit of the motor.

The cylindrical surface of the drum is carefully polished in order to provide better contact with the material being processed.

The material to be separated is fed to a hopper 12, from where it is poured by means of a measuring hopper 9 onto the revolving drum. Revolution of the measuring disk 9 is accomplished from pulley 6 through a crossed drive belt, 7, and pulley 8. To regulate the apportionment of the material, pulleys 6 and 8 are made with steps, so that transferring belt 7 from one step to another can decrease the speed of sprocket wheel 9. At a certain distance from the drum, on the path of its revolution, a band-shaped electrode 11 is mounted; its position is subject to regulation.

Drum 5 and the metallic body of the whole separator are grounded and a high tension negative is led up to them. The positive high tension is fed to electrode 11.

The apparatus for testing cable AKI-50 serves as the source of high tension. This is a fairly compact, small-sized and safe instrument to use which produces regulated, direct, high tension up to 50,000 volts.

The process of separation, as it is possible to see in the diagram, proceeds in the following manner. Particles of the separating material drop from hopper 10 through the measuring hopper 9 onto the drum 5, are charged, negatively, and are deflected to the positive electrode 11 mounted in the corresponding fashion. Thus, they have trajectories which differ for particles having different unit charges.

By changing the position of the separator sheet 12 and the intensity of the electrostatic field it is possible to obtain optimal conditions for separating the desired mineral fractions which fall in the first receiver while the rest of the particles drop in the second and third receivers. A special brush (not shown in the diagram) sweeps off the particles that adhere to the drum.

We constructed the electrostatic separator for the purpose of obtaining clean mica fractions.

Micas are repelled from the separator drum to a considerably greater distance than quartz and feldspar. The flat shape of mica flakes contributes in considerable degree to this effect by providing a large area for their contact with the drum.

By building up the tension of the electrodes to 2,000 to 3,000 volts and passing the fraction being enriched through the separator 2 to 3 times, it is possible to obtain from samples of granite a practically 100 percent mica fraction, free of concretions and consisting of muscovite and biotite. The latter are easily separated from each other in the electromagnetic separator.

The tests we have made dhow that the electrostatic separator may be used successfully also for separating glauconite from sedimentary rock particles.

Having an average productive capacity of

about 8 kg hr of raw materials, the separator uses a total of 0.6 kwt (both in the supply of the AKI and for the motor).

The method we use in working up the rock specimens is as follows. After the disintegration of the rock and the screen analysis, the fractions are separated in sizes 1 to 0.5 and 0.5 to 0.25 mm. Then these fractions are washed to remove dust and are dried in a thermostat. Next the fractions are passed first through the electrostatic and then through the electromagnetic separator.

The degree of cleanliness of the selected mineral fractions is checked with binocular microscope.

Practical use of both separators has shown that with their help not only can we obtain monomineral fractions from crushed rocks, but in some cases, we are able to separate minerals of various generations from these fractions. Thus, for example, we succeeded in separating two varieties of biotite, sharply different in their electromagnetic properties, from a biotitic concentrate of one of the monzonites. The latter circumstance is explained by the fact that granules of magnetite are present in considerable quantity in highly magnetic biotite; they are separated in the process of dissociation of the first generation of biotite.

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NECROLOGY

LOSSES OF SCIENCE

Professor of Leningrad University, Osip Markovich Ansheles, one of the greatest of Soviet crystallographers, died on July 23, 1957. He was born October 17 (5), 1885 in Penza. In 1913 he graduated from the Physico-Mathematical Faculty of Petersburg University with specialties in geology and mineralogy. Two years later O.M. Ansheles was invited by E.S. Fedorov to the Mining Institute for teaching the Fedorov crystal chemical analysis, and from 1924 he headed the chair in crystallography created for him at Leningrad University. Working fruitfully on a wide complex of questions in crystallography, O.M. Ansheles attained the greatest success in the study of the optical properties of minerals, in the field of crystallography, and in the development of methods for the goniometric study of crystals. He wrote more than 70 scientific works including a number of long monographs and textbooks.

Candidate of geological-mineralogical sciences Alexander Fedorovich Sosedko died August 27, 1957. See necrology in "Izvestiya, U.S.S.R. Academy of Sciences, Geologic Series," no. 4, 1958, p. 118.

Professor Sergei Alekrandrovich Yakovlev, great geologist and student of the Quaternary, died October 16, 1957. He was born October 7 (September 25), 1878, graduated from Petersburg University in 1903 and began work there as custodian of the Geological Cabinet. Next he taught geology and mineralogy at the Forestry Institute (S.M. Kirov Technical Forestry Academy) and in recent years headed the Quaternary Department of the VSEGEI (All-Union Geologic Institute). The principal works of S.A. Yakovlev relate to Quaternary geology. He conducted the great scientific organizational work in preparing an international map of Quaternary deposits of Europe. Published twice (1932, 1950) under his editorship, the map of the Quaternary deposits in European U.S.S.R. received general recognition. The detailed stratigraphic diagram of the subdivisions of the Quaternary Period for the U.S.S.R. that he worked out is of great scientific value. The works of S.A. Yakovlev in petrography, hydrogeology, and soil science are well known. He devoted much attention to pedagogical work. His text book of general geology has passed through nine editions. S.A. Yakovlev was awarded the order of Lenin.

Mariya Aleksandrovna Bolkhovitinova, professor of paleontology of the Moscow Geological Prospecting Institute, died October 20, 1957. She was born June 8 (May 27) 1877 in Moscow, in 1907 she entered the University of Shanyavsk as an auditor in the geological department, and in 1911 began work there as an assistant. In 1916 she completed the Higher Women's Courses. After the October Revolution, M.A. Bolkhovitinova began to teach paleontology at Moscow University (in the chair of M.V. Pavlova) and at the Moscow Mining Academy. She attained the rank of lecturer in 1930 and five years later became professor at MGRI (Moscow Geol. Prospecting Inst.). The principal works of M.A. Bolkhovitinova were devoted to the paleontology and stratigraphy of the Carboniferous system of the Moscow area and northeast Kazakhstan. Her work on brachiopods, bryozoa, and sponges is of very great interest. During her long years of pedagogical activity, M.A. Bolkhovitinova trained a great number of geologists and paleontologists. More than 20 published works are from her pen.

Hydrogeologist, member of the Communist Party of the Soviet Union, Mikhail Il'ich Kof, candidate of technical sciences, died November 2, 1957. He was born June 10, 1902 in Kherson; he graduated from the hydrogeological department of the Moscow Hydrometeorological Institute in 1935. Working in Special Geology in the Moscow Geological Trust and other institutions, M.I. Kof was engaged chiefly in regional studies and prepared a series of hydrogeological maps. In the period 1942-1946 he was in the ranks of the Soviet Army and worked on the hydrogeological conditions of large

constructions on the Volga and the groundwater of the Moscow synclinorium. Since 1951, he was senior co-worker of the F.P. Savarenskiy Laboratory of Hydrogeologic Problems, U.S.S.R. Academy of Sciences. M.I. Kof was awarded the order of the "Red Star" and a number of medals.

Senior scientific co-worker of the Institute of Geology of Mine Deposits, Mineralogy, Petrography, and Geochemistry, U.S.S.R. Academy of Sciences, doctor of chemical sciences, Yakov Iosifovich Ol'shanskiy, died January 6, 1958. He was born April 12, 1912 in Lugansk (Voroshilovgrad) and graduated from Novocherkassk Industrial Institute in 1934. In 1937 he began teaching physical chemistry in that same Institute. In the period 1939-1946, Ya. I. Ol'shanskiy served in the Soviet Army and then began work at the Institute of Geological Sciences AN SSSR; he was engaged in problems of experimental mineralogy and petrography. His principal works were devoted to the study of the conditions of physical-chemical equilibrium of various sulfide-silicate systems. He was the author of 32 scientific works.

V.V. Tikhomirov, S.P. Volkova

REGINALD ALDWORTH DALY

Reginald Aldworth Daly -- one of the greatest contemporary petrologists, a scientist of many, broad, interests, author of numerous original investigations in the fields of physical geology, petrology, geophysics, and volcanology, died in Cambridge (USA) on September 14, 1957.

Using a broad approach to the treatment of each question, striving to use and link up the results obtained in related fields of science were as characteristic of Daly as of an explorer; it is possible to group his work in several distinct divisions; nevertheless, work in petrology and geophysics occupy the principal position in Daly's scientific creative genius.

The result of his five-year studies of the geology of the North American Cordillera in the border area of the USA and Canada is a large two-volume work "Geological Structure of the North American Cordillera Along the 45th Parallel (1912); he also prepared a number of separate articles, in particular on the Okanogan batholith (1906), on the region of Lake Musvak (1912), on the geologic structure of the region along the Canadian Railroad (1915). The complex and multifarious magmatic manifestations in the regions he studied attracted his special attention, and in these, as in future works, Daly was already considering a number of general questions of petrography: the classification of rocks, deep (abyssmal) injections, the causes of magmatic activity, the classification of the forms of intrusive bodies, the problem of amplitude (for the solution of which he advanced his well-known theory of the collapse of the cover), the origin of alkaline rocks, and a number of other problems.

In this same group of regional geologic or petrographic studies may be classed Daly's studies on the geology and petrography of the islands of St. Helena and Vozneseniye (1925, 1927), the Bushwell massif (1924, 1928), the structures of Vredfort in South Africa (1947), and of some others in which the chief attention was paid to the more general problems in the given regions.

In an interesting work on the Vredford ring structure (1947), Daly's attention was attracted by interesting geologic facts which could not be explained in the framework of the usual geologic ideas and demanded the application of hypotheses from other fields of knowledge. By partly developing certain ideas expressed earlier, Daly proposed another explanation of the Vredford structure. In his opinion, the fall of a huge meteorite had occurred here; the bursting of this meteorite sharply disturbed the existing gravitational regime and caused the rise to the surface of masses initially located at great depth. This displacement caused the movement to the surface of the ancient granite basement and the formation of an anticline. This hypothesis is supported, in Daly's opinion, by the existence of intrusions of gabbroid rocks, occurring due to the injection of basic magma which penetrated the cracks of the fissure which had developed in the granite anticline.

In his studies Daly repeatedly returned to the problem of the connection between the movements of the ocean bottom, the development of coral reefs, the epochs of glaciation, and also to the questions of the primary chemical composition of the waters of the ancient ocean. Daly explained the absence of life in the Precambrian ocean by the insufficient quantity of calcium salts in the sea water of that time. Daly's theory also explained certain questions of the evolution of life on Earth in ancient time and the genesis of dolomitic rocks.

In a series of works in 1916, 1929, 1934, and 1948, Daly reviewed the relation between the phases of glaciation in the

Pleistocene and the history of the change of level of the Pacific Ocean in connection with a supposed mechanism of the deformation of the Earth as a whole, based on geophysical theories. In 1948, he summarized all the new data and came to the conclusion that, in general, they supported his ideas expressed earlier. In this group should be classed Daly's works devoted to coral reefs and atolls, and also to work on submarine "canyons" (1936), the origin of which Daly connected not with the flooding of river valleys, as did the majority of investigators, but with the activity of ocean currents in the coastal belt.

Throughout the whole course of his scientific activity, Daly gave precedence to questions of petrology and geophysics in connection with intrusive (plutonic) and volcanic processes, and also the composition of the Earth's crust and plutonic geospheres. He was also the author of one of the most famous theories of petrology -- the assimilation of alkaline rocks as a result of the interaction of alkaline magma with carbonaceous rocks. In defense of his theory and as proof against the serious objections that quickly arose at the time of its appearance, he offered broad factual evidence on the majority (about 150) of alkaline rock deposits known at that time in various parts of the world, and also a number of general considerations on the chemistry of alkaline rocks, and their typical associations with other igneous rocks. In stating his theory, Daly clearly presented its weak sides and persistently pointed out that he did not have physico-chemical proofs of laboratory experiment. By its emergence, this theory, although false in considerable measure, nevertheless played a large positive role in directing petrographers to a deeper study of these complex and important problems of petrogenesis.

In another of his works on petrography, Sills and Laccoliths as Illustrations of Petrogenesis (1918), Daly expressed interesting ideas on the fact that these small intrusive bodies, when not of uniform structure, reflect in miniature the most general laws of petrogenesis. The small dimensions of these bodies and the consequent small heat storage did not contribute to increased assimilation of lateral rocks, and in this connection the chief factor of petrogenesis was differentiation.

In these works Daly gradually approached the creation of his synthetic-differentiation theory of petrogenesis, developing it simultaneously and in parallel with F. Yu. Levinson-Lessing. A further, and at the same time most complete and systematic, exposition of Daly's views in the field of petrogenesis and other questions of petrology was his well-known book, Igneous Rocks and the Depths of the Earth (1914), which remains today one of the basic monographs in the field of theoretical petrology. Based on a vast amount of factual material, largely personal observations, this book is a synthesis of the petrographic science of that time; it gives the general theory of magmatic processes and points the way to working out individual problems. In his views, most systematically expressed in this work, Daly is a proponent of the independent existence of two magmas in the Earth's crust -- acid and basic, departing in this respect from the most widespread views of that time held by Bowen and other petrographers.

Daly returned to the development of individual aspects of the general problems of petrogenesis in his later works. Such were his studies of the Bushveld complex in South Africa (1924, 1928) in which he evolved the idea of the origin of the complex as the result of processes of crystallizing differentiation, occurring on the spot. In a work devoted to the problem of granitization (1949), Daly subjected to critical review the studies of the "metasomaticists" -- Baklund, Reed, Rubo, and others. In Daly's opinion, granitizing emanations could not rise from the lower-lying peridotite and basalt layers, but came only from the upper granite, sial layer, the existence of which the "granitizationists" ignore. The huge masses of basalt magma, said Daly, lying at the base of the Canadian shield and platform since Paleozoic times, should have exuded since that time great masses of acid differentiates, nevertheless this has not been observed.

Numerous other works of Daly take much this same direction, as does his theory of volcanic processes, in the working out of which he discussed the general problems of petrology and geophysics. In a work of 1911 he reviewed the question of the nature and causes of volcanic activity. In it, he underlined the connection of volcanic processes near the surface with the processes of a plutonic "abyssmal injection" and made a division of volcanic gases into magmatic and phreatic types. The source of heat in volcanic processes is connected not only with the initial heat of the plutonic cell, but also with additional heat developing during exothermal reactions. Prolonged activity of the central cell is explained as the result of two-phase convection occurring in the central volcanic canal.

Daly turned also to questions of the theory of volcanism in his works devoted to the geology and petrology of the volcanic islands of St. Helena, Vozneseniye, and Hawaii (1944). Examining the reasons for

e varied composition of Hawaiian lavas, aly came to the conclusion that differentia- on in the given case was determined by the pecific weight and the character of diffusion f the volatile components in the magma within the cell). The acid components of the eries (trachytes) emerged, in his opinion, s a result of crystallized differentiation eing exposed to further "flotation." Alkaline d melilitic basalts arose during the inter- ction of basalt magma with carbonaceous ocks. All the conjectured processes of dif- erentiation, refusion and others, took place hiefly in the abysses, which next gave the aterial for vertical injections. In Daly's pinion, there is no special difference be- ween the "continental" and "oceanic" basalts.

The concrete application of geophysical uestions to the solution of problems of etrology occupied an important place in aly's works, especially in his later years. uch were his works on the influence of the owing of the Earth's rotation on geologic rocesses (1943), on the structure of the uter envelopes of the Earth's crust (1928), the structure of the Earth's crust under e oceans (1943), on the problems of isos- sy (1926, 1929). In his book Strength and ructure of the Earth (1940), Daly analyzed e geologic and geophysical data amassed to that time from the point of view of e theory of isostasy, examining them in ccession for all continents and oceans. He me to the conclusion that this theory rees in general with the idea of the exist- ce of an elastic envelope -- an astheno- here with a depth of 80 to 100 km, eakened in stability (with lower pressures) comparison with the upper and lower yers of the lithosphere.

In this same general range of problems ay be classed Daly's work on the history the geologic development of the ocean ttom depressions -- The Floor of the cean -- New Light on an Old Secret (1942), ken from a course of popular lectures. In is book are formulated important conclu- ons about the wide extent of submarine lcanism, the existence of submarine ountain ranges comparable to the highest ountain ridges on the surface of the conti- nts. On the basis of seismic data, Daly nsiders that in the expanses of the ocean oor (outside shelf areas) there are subma- ne masses having the properties of conti- ntal blocks. The same data, in his opinion, nfirmed the existence of an asthenosphere depths of an order of 80 km, probably der conditions of high temperature.

Daly returns to the question of the nature the asthenosphere in another of his books 946). Using some new geophysical data, recognizes as more correct the idea of the state of the asthenosphere as two-phase rather than glassy, as he had thought earlier. In presenting this new hypothesis, Daly indicates at the same time his lack of direct proofs and the need for setting up experiments to clarify the influence of high temperatures on the compressibility of liquid, glassy, and crystal substances.

To this same group also belongs Daly's article on the origin of continents and the "continental hemisphere" (1951). Here he stressed the significance of the horizontal differentiation of the undercrust substance in pre-geologic times, side by side with its vertical differentiation. Daly advanced the idea of the relatively later age of the Atlantic and Indian ocean depressions along with this theory.

For explanation of the structure and conditions of development of the Earth's envelopes (geosphere) Daly also drew on material from the study of the composition of meteorites (1943). Comparison of the composition of various types of meteorites led him to the conclusion that the latter are the remains of a disintegrated planet analogous to the Earth in composition and historical development. The sharp differences in the composition and texture of meteorites of different types -- rock, iron, and others -- are a reflection of sharp differences in the qualitative composition of individual envelopes of the given meteorite, and consequently, also of the Earth.

Even a brief summary of the principal scientific works of R.A. Daly reveals the great variety of his interests, his profound erudition in the field of geologic science, and the boldness of his comparisons and conclusions. These marks of a scholar created for him a wide popularity not only on the American continent, but also beyond its borders. He was a member of the National Academy of Sciences in Washington, an honorary member of a number of academies of science in other countries. He was the editor-in-chief of the well-known publication American Journal of Science.

Notwithstanding that many of the views and theories expressed by Daly in his time were subjected to serious criticism and revision or even complete rejection, Daly's role in the history of the development of the sciences connected with the study of the material composition of the Earth during the first half of our century was extremely significant and universally recognized.

A.P. Lebedev

REVIEWS AND DISCUSSIONS

ON THE ARTICLE OF S.I. BALASANYAN THE ORIGIN OF THE BASIC DIKE ROCKS OF ARMENIA AND THE ADJACENT SECTIONS OF MALYY KAVKAZ[1]

by

G.A. Kazaryan, E.G. Malkhasyan, and Yu. A. Leie

An article by S.I. Balasanyan was published in "Izvestiya, U.S.S.R. Academy of Sciences Geologic Series," No. 7, 1956, entitled The Origin of the basic dike rocks of Armenia and the adjacent sections of Malyy Kavkaz.

In this article, the author writes that all dikes of basic composition in Armenia and the adjacent sections of Malyy Kavkaz developed as the result of activity of an independent magmatic body, not connected with the source of intrusions; the time of this activity, in his opinion "is quite definitely established as Tertiary."

Many geologists have been engaged at various times in the study of veins in connection with research on magmatic rocks; the authors of the present article have also paid due attention to this problem in their work in various regions of Armenia. However, the data we have received not only do not allow us to agree with the conclusions of S.I. Balasanyan, but furnish a basis for considering them exaggerated.

Contrary to the assertion of the author (p. 79) the list of literature he used far from exhausts all the works even on the veined rocks of Armenia, not to mention the Soviet Union as a whole. Moreover, the attached list of sources, for example, the works of A.T. Aslanyan, A.A. Gabrielyan, transactions of the First Conference on Questions of Cosmogeny, has no connection with the problem being discussed (there is not even a reference to this paper in the text); the widely known works on vein formations of M.B. and H.I. Borodayevskiy and O.S. Polkva, and works on Armenia -- of V.N. Kotlyar, G.P. Bagdarsaryan, P.F. Sopko [5], E.G. Malkhasyan [4], and others are missing. The incompleteness of the list of references is, perhaps, reflected in the conclusions of the author.

The main thesis of the article is encompassed in the statement that "the basic dikes broke through after the formation of all intrusives and their veins, i.e., they are of post-upper Eocene age" (p. 80). In confirmation of such a conclusion, the author puts forth the following data:

"1. In some places the basic dikes cut the intrusives.

"2. On the divide line of Gedzhalinskiy Ridge the basic dikes break through brecciated and mylonitized rocks of a granite mass, remaining themselves without any traces of fracture. ...

"3. To the west of the city of Dilizhan hydrothermally altered igneous rocks, affected by the thrust of an intrusive of alkaline composition, are burst open by the basic dikes which, themselves, remain fresh.

"4. To the east of the city of Kirovakan, in the breccias and hydrothermally-altered igneous rocks, we found at a considerable distance from a granitoid intrusive a fresh dike of basic rocks." (p. 80-81).

At first glance it may seem that these astonishing data could confirm the hypotheses being presented. However such an impression is mistaken: the "factual material" with which S.I. Balasanyan is working is far from characteristic of the parts of Malyy Kavkaz adjacent to it; he does not even show the positions of the dike rocks in the

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[1] The editor also received critical remarks on the article of S.I. Balasanyan from K.A. Karamanyan and T.A. Arevshatyan -- co-workers of the Institute of Geological Sciences, Armenian S.S.R.

In view of the fact that their criticism of the views of S.I. Balasanyan do not differ substantially from those published below, the article of G.A. Kazaryan, E.G. Malkhasyan, and Yu. A. Leie, appears in the present number, because it was received by the editor before the article of K.A. Karamanyan and T.A. Arevshatyan.

nsignificantly small sections which were in-
ependently investigated by the author before
ubmitting the article for publication.

Numerous facts speak out against the con-
lusions of S.I. Balasanyan, indicating that
is hypotheses were preconceived, and point
ut the incompleteness and the clearly inac-
urate interpretation of the facts presented;
he author ignores the age of basic dike
ocks and their distribution in the various
eotectonic zones. He takes into considera-
ion only a certain similarity in chemical
omposition and petrochemical peculiarities
for which he prejudicially uses a total of
nly 6 purposefully selected analyses). Dike
ocks of the basic mass, of various ages and
f sharply different geotectonic zones, with
pecific features of geologic development,
agmatism and metallization he combines in
ne group, as of the same age -- Tertiary --
nd having a single source. This might be
rue for a small region in the limits of one
eotectonic zone, but not for all of Armenia
nd the provinces adjacent to it which are
f a rather complicated geologic structure.

It is generally known that numerous dike
ocks of basic composition, particularly
iabase, diabasic porphyry, and gabbro
orphyry of the Alaverdy and Kafan ore-
earing regions of Armenia were intruded
uring Mesozoic time and in the majority
f cases, are of Jurassic age. Besides the
lesozoic and Tertiary (Eocene and Oligo-
ene) intrusive cycles, with which the appro-
riate vein rocks were associated, there is
complex of dike rocks of basic composi-
on connected with the Kongur-Alangez
lateau in southern Armenia, the age of
iese rocks is believed to be Miocene. The
uestion is, could basic dike rocks of the
laverdy or Kafan regions, which do not
ppear beyond the limits of the southern
ikes of Mesozoic age, be intruded simul-
ineously with the dikes which burst through
ie Kongur-Alangez granitoids of Miocene
ge.

In illustration let us cite the character-
stics of the vein complex of the Alaverdy
re-bearing region (after G.A. Kazaryan),
e., the region where S.I. Balasanyan
orked and gathered his "factual" data.

The vein rocks of the Alaverdy region
re subdivided into two groups associated
ith effusive and intrusive cycles of mag-
atism according to their geologic and
ructural position, mineral composition and
emical peculiarities.

The sequence of intrusion of these dikes
as follows: 1) vein rocks of the Mesozoic
fusive cycle: microporphyry, andesite
orphyry, quartz albitophyres, dacite porphyry, potash granite porphyry; 2) vein rocks associated with the Mesozoic intrusive cycle (quartz diorite and granodiorite); vein rocks of the first stage: aplite and pegmatite; vein rocks of the second stage: dolerite, gabbro porphyry, dioritic porphyry, potash microgranite, pyrite veins (hydrothermal origin); 3) vein rocks associated with the Tertiary intrusive cycle (quartz diorite, granodiorite, quartz diorite porphyry and others); vein rocks of the first stage: aplite, granite porphyry; vein rocks of the second stage: dolerite, plagioclase porphyry, diorite porphyry, granodiorite porphyry, and metallic and barite veins of hydrothermal origin.

This division of the dikes into groups associated with intrusive and effusive cycles and their various ages were discovered in the other regions mentioned by S.I. Balasanyan, where the connection of the dikes with Jurassic effusives was fully established and criteria determined by similar, but different, dikes of various ages associated with various cycles of magmatic activity.

The assertion of the author that "for Armenia the Tertiary age of the basic dikes is fully established" (p. 84) has no basis whatsoever. A number of investigators (S.A. Movsesyan, V.N. Kotlyar, G.P. Bagdasaryan, E.G. Malkhasyan, P.F. Sopko, and others) associate the vein with granitoid rocks and even find it possible to link up different varieties of dikes genetically with quite distinct phases in the intrusive cycle. Their data stem from thorough analysis of factual material gathered in the regions of intensive multiple manifestations of magmatism in the Armenian SSR.

Thus, V.N. Kotlyar described the close genetic connection among all rocks of an intrusive complex in 1930 on the model of the Gyumushkhan intrusive complex, including monzonite, essexite, granophyre, and anorthosite. The presence of similar and extremely characteristic associations of minerals in different varieties of these rocks gives evidence of their origin from one magmatic nucleus. The intrusion of the two last varieties, apparently, came from the deepest part of the magma reservoir.

Denying the genetic connection of the basic dikes with granitoids, S.I. Balasanyan alluded to the fact that there is "quite a considerable age limit separating the basic dikes from the intrusives and their leucocratic derivatives (Northern Armenia)" (p. 82). A similar conclusion does not follow from the study of the dike rocks of Northern Armenia; it can be extended only to part of the Alaverdy region, i.e., a comparatively small section, but even here, obviously, a number of dikes were not taken

into account by the author.

A certain lag in the intrusion of basic dikes and other vein rocks of the second stage refutes such hypotheses in certain ore-bearing regions (Alaverdy, Shamlug, Akhtala, Kafan, and others), in reviewing the question of the interrelationship of the processes of hydrothermal activity and dike formation of the second stage; and the author's reference to pre-mineral dikes, after S.S. Mkrtchyan and T.A. Arevshatyan (p. 81), also contradicts his conclusions.

S.I. Balasanyan did not sufficiently discuss the question of the interrelationship of the dikes of basic composition with hydrothermal activity. He takes under consideration only the first stage of hydrothermal activity, indirectly connected with exposed granitoid intrusives, and finds that the dikes of basic composition are retarded and, consequently, have no connection at all with these granitoids. The author does not take into consideration the fact (and herein lies his error) that the chalcopyrite mineralization in Northern Armenia and the copper molybdenum in Southern Armenia was caused by hydrothermal activity at the end of the formation of the vein rocks of the second stage, and is the final stage in the Mesozoic and Tertiary intrusive cycles (Alaverdy and Kafan and Kongur-Alangez).

It is worthwhile to mention that in the Alaverdy ore-bearing area the dikes of the second stage of the Mesozoic intrusive cycle, including the Shamlug chalcopyrite veins, have a uniform structural position and that the contact aureoles of the nearly latitudinal basic dikes indicate extremely high concentrations of copper, zinc, iron, and cobalt. Thus, it may be stated with some credibility that the vein rocks of the second stage and the hydrothermal veins subsequent to them are probably derivatives of a deep-seated magma. The multiple stages of the formation of the vein complexes in Armenia are apparently the result of the activity of separate magma reservoirs performing at different times.

Finally, it is not quite clear of what "considerable distance" between the dikes and intrusives the author is speaking, whether from the large Kokhp-Shnokh intrusive mass to Lal'var mountain not over 20 km, or from Akhpatsk -- 5 to 7 km, because the dike rocks of basic composition are localized near these intrusives and are attracted to it.

We find the data of M.B. Borodayevskaya quite convincing [1]; she divides the dikes into three genetic types: 1) those connected with intensive volcanic activity in the early stages of development of geosynclines, 2) derivatives of large intrusions during the middle stages of geosynclinal development, 3) the late stages of geosynclinal development. These last types, in her opinion, are independent formations arising in the semiplatform period of development of the region and are not connected with either the large intrusions, nor with effusive activity.

The new data on the origin of dike rocks, resulting from the analysis of a great deal of factual material, is known through the works of V.S. Koptev-Dvornikov, O.S. Polkva, M.A. Pavorskaya, and a number of other investigators who to a certain degree supplement and modify the hypotheses of M.B. Borodayevskaya. The views of the authors mentioned are receiving the approval of many contemporary geologists and are attracting a good deal of attention.

Without denying the possibility of the existence of independent vein complexes (in the concept of M.B. Borodayevskaya), i.e., dikes, completely unconnected with large intrusions, we, nevertheless, claim that they could not appear in the limits of Malyy Kavkaz, because similar complexes develop only in the late and final stages of development of active belts, and the folded zone of Malyy Kavkaz has experienced only the first two stages of development.

It should also be pointed out that S.I. Balasanyan, in trying to substantiate his views, incorrectly uses the data of other investigators who do not fit into the framework of his hypotheses. Thus, citing the work of V.S. Koptev-Dvornikov, he speaks of the lack of genetic connection between melanocratic veins and intrusives (p. 84), while V.S. Koptev-Dvornikov, as is known, not only does not deny such a connection, but points out with ample clarity that "the dependence of these rocks on the composition of the intrusions is determined by a number of circumstances" ([2], p. 75). V.S. Koptev-Dvornikov also develops this same position in his subsequent works. He relates granite-porphyry, diorite-porphyrite, dolerite, gabbro porphyry, and lamprophyre to the indicated rocks of the second stage. Lamprophyres, in his opinion, are related to ordinary diorite porphyry and do not represent an independent series of rocks or derivatives of a special lamprophyritic magma. He remarks further: "If vein rocks of the first stage are spatially connected with the intrusive bodies themselves, the vein rocks of the second stage issue, probably, from deep-seated reservoirs ([3], p. 137).

An analogous example may be found in the fact that the dike rocks of basic composition described in the works of A.L.

odin, I.N. Sitkovskiy, and other investi-
ators of Armenia, to which S.I. Belasanyan
efers, are viewed as genetically related to
e intrusives they studied, although in indi-
idual cases these dikes do not border the
ntrusives closely enough. S.I. Balasanyan,
iting these investigators, and being unfamil-
ar with the magmatic activity in the regions
f their works, with a single flourish of the
en tears away the dike rocks of basic com-
osition from the intrusives to strengthen
is opinion.

The inadequate attention the author pays
the working up and generalization of the
ata from the literature on the question
nder discussion is also shown by the fact
at the one outcrop of ultrabasic rocks
nown in the Alaverdi region, which O.S.
tepanyan defined as picrite, and N.A.
orozov as picrite with plagioclase, S.I.
alasanyan takes as different outcrops and
emonstrates how very close the formations
re to each other, drawing from this false
nd unnecessary conclusions (p. 82). This
explained by the fact that, at the time of
is field works in this region S.I. Balasan-
an did not personally observe this interest-
g dike.

In S.I. Balasanyan's paper, there are also
number of other inaccuracies, but for
ck of space we cannot take them up here.

And so, S.I. Balasanyan's attempt to give
new explanation of the origin of the dike
ocks of basic composition of Armenia and
e adjacent sections of Malyy Kavkaz can-
ot be considered a success.

In conclusion, it should be mentioned
at in our article we did not undertake to
xamine the origin of the basic dike rocks
f Armenia and the adjacent parts of Malyy
avkaz. Such an important study merits a
pecial article. Much factual material on
e problem is in the stage of preparation,
ut it is already possible to assert now
ith sufficient confidence that the dikes of
asic composition are genetically related to
e processes of intrusive and effusive mag-
atic activity.

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LETTER TO THE EDITOR

Dear Editor:

In regard to my discussion with Yu. A. Kuznetsov on Iron-mining and genetic types of intrusions (Izvestiya, U.S.S.R. Academy of Sciences, Geologic Series, no. 2, 1955, and no. 7, 1957) I submit the following:

Yu. A. Kuznetsov explains the appearance of my remarks as a misunderstanding caused by the fact that in his article the ratio FeO:MgO is given as a ratio of molecular numbers, and in my Critical Remarks (Izvestiya, U.S.S.R. Academy of Sciences, Geologic Series, no. 8, 1956) -- as if in weight percent. That, of course, is not the

case. I, like Yu. A. Kuznetsov, was using molecular numbers.

I give two examples. The chemical analyses of syenites were taken from the account of Yu. A. Asanov for 1936, where the molecular numbers are given in parentheses. 1) Test 13, borehole 25. Fe_2O_3--3.14 (021), FeO--0.79 (011), MgO--0.22 (005); let us convert Fe_2O_3 to FeO and then we get FeO--53 (21 · 2+11); dividing 53 by 5 we get FeO:MgO--10.6. 2) Test 28, borehole 34. Fe_2O_3--1.13 (007), FeO--1.89 (026), MgO--2.53 (062); FeO:MgO--0.6.

From this follows the conclusion that the given data are unreliable for separating syenites into gamma- and beta-syenites.

V.G. Korel'

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